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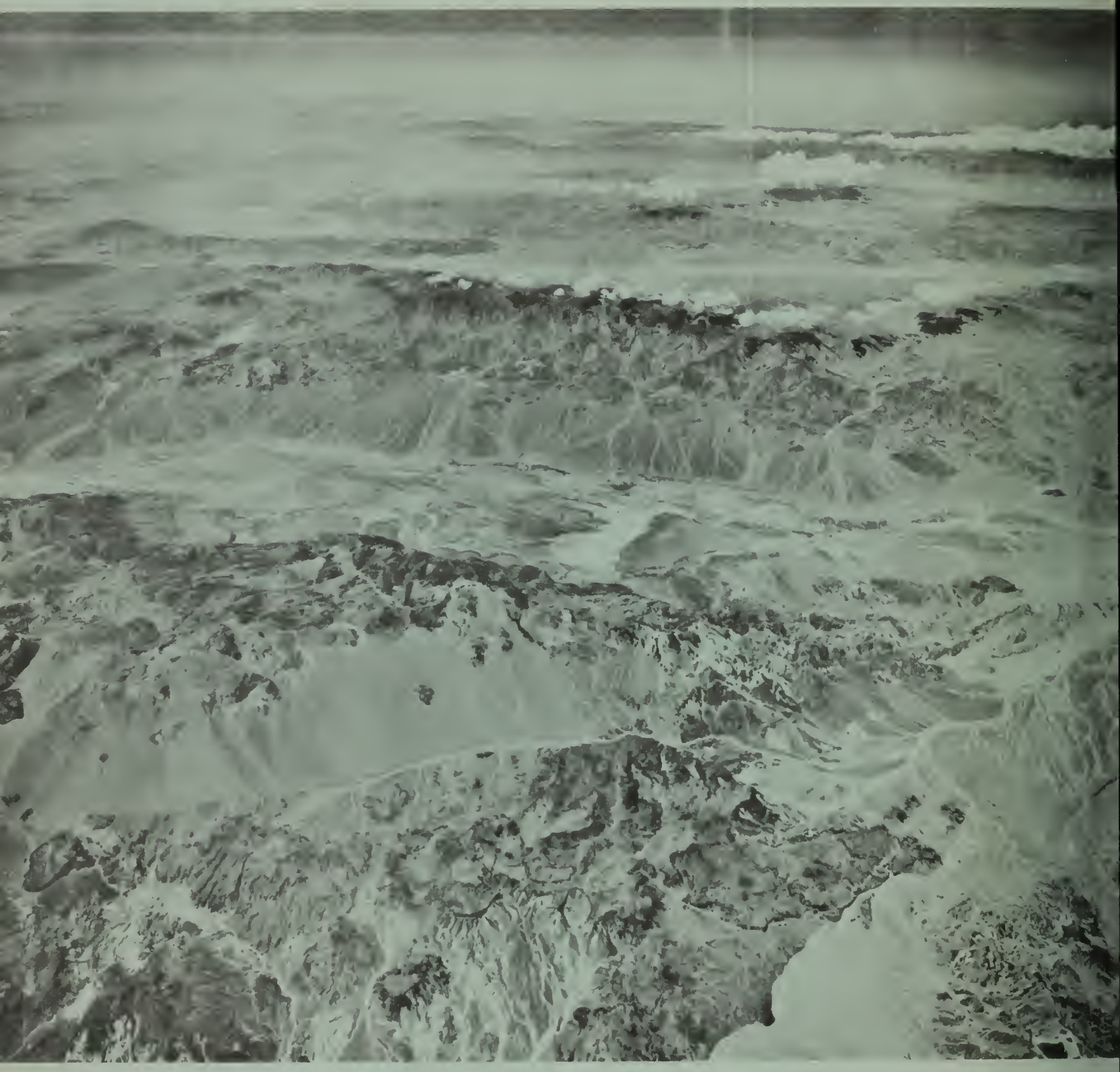

Inside front cover. High-altitude oblique aerial photograph oriented approximately west-southwestward across Death Valley. The valley between Furnace Creek (right) and Smith Mountain (left) extends across the central part of the photograph. The Black Mountains, Greenwater Valley (and Furnace Creek Wash) and the northern part of the Greenwater Range are towards the observer. A small segment of the Funeral Mountains farms right foreground. Badwater is near white patch in center of photograph. The Panamint Range on west side of Death Valley has a line of scattered clouds along its crest.

Patch of clouds in upper right is approximately over Mt. Whitney in the Sierra Nevada. *Photo U.S. Air Force 018R 194, 10 July 1968; courtesy of the U.S. Geologicol Survey.*

Front cover. High-altitude oblique aerial photograph oriented west-southwestward across north-central Death Valley. The crest of part of the Panamint Range lies beneath the string of clouds just beyond central Death Valley. Tucki Mountain is directly under small cloud near center of photograph. Small dark patch on right edge of symmetrical alluvial fan barely left of center of photagraph is Furnace Creek Ranch. Mass of clouds at upper right are near Mt. Whitney in the Sierra Nevada. Funeral Range occupies foreground. *Photo U.S. Air Force 018R 191, 10 July 1968; courtesy of the U.S. Geological Survey.*

SPECIAL REPORT 106

GEOLOGIC FEATURES
DEATH VALLEY, CALIFORNIA

Revised Papers Selected From a Guidebook on the
Death Valley Region, California and Nevada
Published in 1974

Edited by

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1976

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High-altitude vertical aerial photograph of floor of north-central Death Valley. Top of photograph is to north. Symmetrical alluvial fan at left upper edge of photograph is at mouth of Furnace Creek Wash. Black Mountains on right and east margin of Panamint Range on lower left. Badwater is on right edge of prominent white patch in right-central part of photograph. East-dipping Paleozoic strata form v's in part of Panamint Range. Hanaupah fault truncates fans from Panamint Range (lower center). Very young faults bound steep smooth west side of Black Mountains (right). *Photo U.S. Air Force 314V 192, 6 September 1968; courtesy of the U.S. Geological Survey.*

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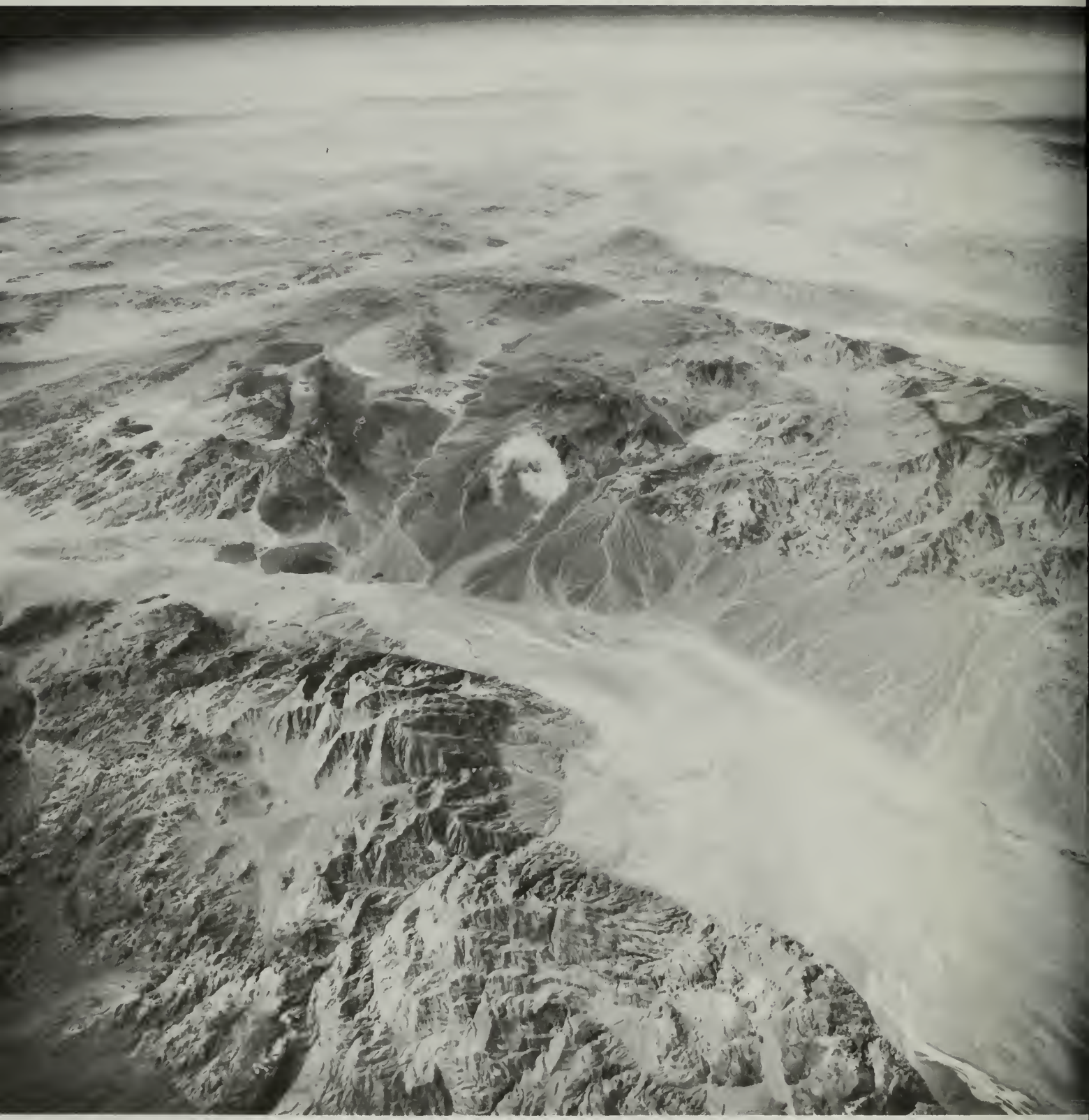

High-altitude oblique aeriol photograph to southwest ocross south-centrol Deoth Volley. Block Mountoins in foreground. Owlsheod Mountoins, left center, and sauthern Ponamint Ronge, right center, are on opposite sides of Wingate Wosh. Gorlock fault extends obliquely ocross upper ane-third of photogroph an for side of Owlshead Mountoins. Ronge with bulbous south end (upper right center) is Slote Range. Seorles Lake lies to right af Slote Ronge. San Bernordino and Son Gobriel Mountoins and Sierro Modre near horizon beyand Majave Desert. Sauthern end of Sierro Nevoda in upper right. *Photo U.S. Air Force, 374R 194, 6 September 1968; courtesy of the U.S. Geologicol Survey.*

PREFACE

This collection of papers and maps was published in 1974 as part of a guidebook prepared for a geologic field trip from Las Vegas, Nevada, to Death Valley.

Several of these papers have been corrected and modified since they were published in the guidebook. The guidebook was published by The Death Valley Publishing Company, Shoshone, CA 92384, for Field Trip No. 1 of the 70th annual meeting of the Cordilleran Section, Geological Society of America. The guidebook contains the following additional papers:

Geologic Guide to the Death Valley Region, California and Nevada, *by Bennie W. Troxel.*

Geology of the Spring Mountains, Nevada, *by B. C. Burchfiel, R. J. Fleck, D. T. Secor, R. R. Vincelette, G. A. Davis.*

Turtleback Surfaces of Death Valley Viewed as Phenomena of Extensional Tectonics, *by Lauren A. Wright, James K. Otton, Bennie W. Troxel.*

Geology of the Furnace Creek Borate Area, Death Valley, Inyo County, California, *by James F. McAllister.*

Significance of a Man-made Diversion of Furnace Creek Wash at Zabriskie Point, Death Valley, California, *by Bennie W. Troxel.*

Correlation (chart) of Lower Cambrian and Some Precambrian Strata in the Southern Great Basin, *by J. H. Stewart.*

Bennie W. Troxel
Geologist
California Division of Mines and Geology
March, 1976

High-oltitude oblique oeriol photogroph west-southwestword ocrass north-centrol Deoth Volley token from o point opproximotely over Beotty, Nevodo. Amorgoso Desert in lower left; Bullfrog Hills, lower right; Funerol Mountoins, left center; ond Gropevine Ronge, right center. Death Volley extends ocross center of photogroph. Panomint Ronge lies beyond the volley. Mt. Whitney in Sierro Nevodo lies approximately beneoth lorge cloud in upper right. *Photo U.S. Air Force 018R 188, 10 July 1968; courtesy of the U.S. Geological Survey.*

Precambrian Sedimentary Environments of the Death Valley Region, Eastern California

L. A. Wright,[1] B. W. Troxel,[2] E. G. Williams,[1] M. T. Roberts,[1] P. E. Diehl[1]

INTRODUCTION

Noteworthy among the geologic features of the Death Valley region of eastern California is an extensively exposed accumulation of sedimentary rock and sill-forming diabase, Precambrian in age, as much as 5,500 m thick, and essentially unmetamorphosed. Although these rocks have yet to yield a reliable radiometric date, the oldest probably were deposited about 1.4 b.y. ago, as they rest with profound unconformity upon a crystalline complex from which 1.7-b.y. dates have been obtained, and they have been intruded by diabase sills and dikes that seem best correlated with similarly disposed diabase bodies, of central and southern Arizona, that are about 1.2 b.y. old. The youngest Precambrian strata of the Death Valley region conformably underlie strata that contain Early Cambrian fossils.

This Precambrian section (Fig. 1) has long been recognized as composed of (1) a lower succession, named the Pahrump Group and referred to by some as "Beltian" in age, and (2) an upper succession divided into several formations (Fig. 1) comparable in stratigraphic position with the Windermere Group of the more northerly parts of the North American Cordillera.

The two successions were originally defined as separated by an angular unconformity beneath the Noonday Dolomite (Noble, 1934; Hazzard, 1937; Hewett, 1956). The unconformity causes the Noonday to rest variously upon each of three formations of the Pahrump Group and upon the older crystalline complex.

Most workers in the region have assumed that somewhere in this impressive accumulation of sedimentary units and sills is recorded a change in provenance from one that was initiated with the beginning of Pahrump sedimentation and has remained little understood, to another that constitutes the provenance of the Cordilleran miogeosyncline. The sub-Noonday unconformity has frequently been inferred to be the principal indicator of this change, so that published stratigraphic cross sections of the southern part of the Cordilleran miogeosyncline consistently show the Noonday as the lowest formation in it.

Not everyone has accepted this interpretation, however, partly because carbonate rock rarely forms the initial deposit in a developing miogeosynclinal environment. Noble (1941, p. 592) observed that the entire Johnnie Formation, which overlies the Noonday, is lithologically more similar to parts of the Pahrump Group than to any of the younger formations, and on many occasions, he informally expressed the belief that the deposition of the enormous volume of detrital silica preserved in the Stirling Quartzite marked the initiation of the miogeosyncline. Stewart (1972), on the other hand, has suggested that the change in provenance is recorded in the pre-Noonday conglomeratic units, commonly diamictite, of the Kingston Peak Formation.

We will briefly summarize data, much of it recently acquired and described more completely elsewhere in this volume, that relates to the nature of sedimentary environments that existed before the beginning of the Cordilleran miogeosyncline. These data indicate that the occurrences of the Pahrump Group in the southern Death Valley region were deposited within a long-continuing west-northwest–trending trough or basin (Figs. 2 and 3) about 50 km wide and at least 140 km long, with bordering platform areas that at times were inundated by shallow seas and at other times stood well above sea level to become source areas of basinal clastic sediments. The structural and stratigraphic characteristics of this feature suggest that it is an aulacogen precisely in the sense that aulacogens are defined and characterized in a recent review by Hoffman and others (1974). As indicated by them, aulacogens were recognized by the Russian geologist Shatiski and consist of "long-lived, deeply subsiding troughs, at times fault-bounded, that extend at high angles from geosynclines far into adjacent foreland platforms." The trough that controlled Pahrump and subsequent sedimentation is named by us the Amargosa aulacogen. Its bordering, generally high areas to the north and south we designate as the Nopah (Wright and Troxel, 1967) and the Mojave uplands, respectively (Fig. 3B).

As the Noonday Dolomite passes abruptly southward into time-equivalent basinal deposits that are approximately coextensive with basinal units of the underlying Pahrump Group, we view the Noonday and its equivalent units as deposited within the provenance of the Amargosa aulacogen rather than within the Cordilleran miogeosyncline. The sub-Noonday unconformity we attribute to strong post-Pahrump emergence along the northern margin of the aulacogen. Thus interpreted, the Noonday Dolomite marks a carbonate platform developed on the deeply eroded edge of a trough.

We also observe evidence that the overlying Johnnie Formation and Stirling Quartzite were deposited on a south to south-southwesterly paleoslope, tectonically perpetuated since the beginning of Pahrump time, and that the northwesterly paleoslope, which seems to have characterized the part of the miogeosyncline that occupied the site of the southern Death Valley region, became firmly established with the deposition of the Wood Canyon Formation. The uplift of the Nopah upland apparently recurred briefly and finally to produce a southwesterly paleoslope upon which were deposited arkosic strata in the lower part of the middle members of the Wood Canyon (Diehl, this volume).

The existence of the depositional trough and the west-northwest orientation of its axis is indicated in numerous stratigraphic and sedimentologic features of the three Pahrump formations and the Noonday. Isopachous lines drawn on these formations collectively, on each formation, and on most individual members display a troughlike configuration. With respect to this configuration, deep-water facies, where present in the section, occupy the center of the trough; some carbonate units intertongue with siliceous clastic strata near the margins; clastic units of various types fine troughward or from one side of the trough to the other; and various current direction features indicate paleoslope directions consistent with a west-northwest–trending trough. Changes in facies and thickness are characteristically abrupt and commonly occur on either side of mappable faults or monoclines. The formation of the trough is thus largely attributable to vertical

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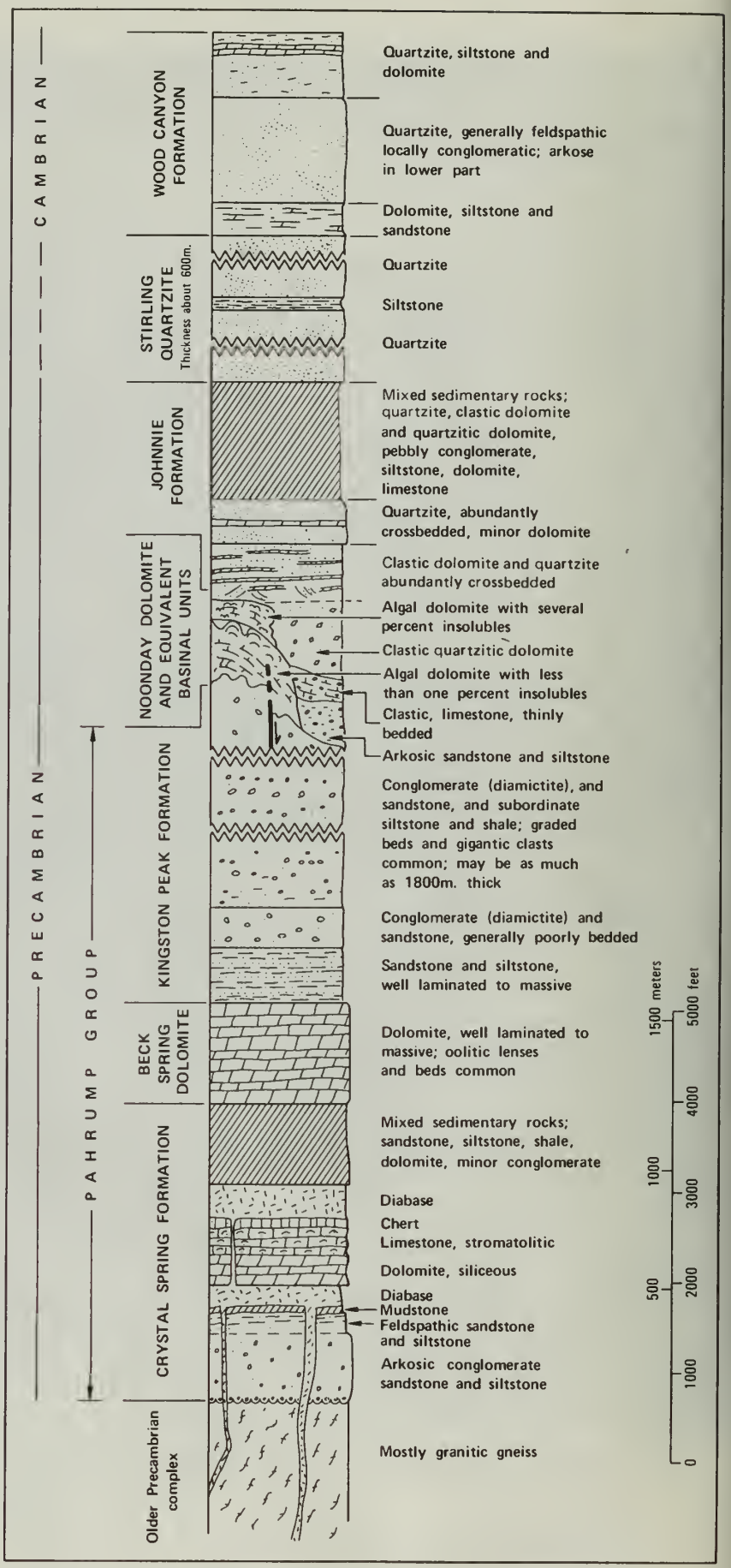

Figure 1. Generalized columnar section of Precambrian to Lower Cambrian strata, Death Valley region.

movements on fault-bounded blocks. The lower part of the Kingston Peak Formation, on the other hand, probably was deposited while the earlier underlying strata were being folded into a broad downwarp.

NOPAH UPLAND

The northerly margin of the trough was first detected in the reconstruction of paleogeologic contacts on the sub-Noonday surface (Wright and Troxel, 1967; Fig. 3C, this paper). Their configuration indicates that, prior to the deposition of the Noonday Dolomite, an uplifted positive area of crystalline rocks extended from the present Black Mountains for at least 100 km eastward and was flanked southward by progressively younger units of the Pahrump Group. Additional evidence for the uplift and for the conclusion that the two lower formations of the Pahrump Group, namely the Crystal Spring Formation and the Beck Spring Dolomite, once extended northward into the area of crystalline rocks on the paleogeologic map and were eroded away in pre-Noonday time (Wright and Troxel, 1967; Troxel, 1967), is contained in the observations that Crystal Spring and Beck Spring debris is abundant in the conglomeratic units of the Kingston Peak Formation and that the debris was transported southward.

Other observations, made subsequently, indicate that this northerly positive area and the south to south-southwesterly paleoslope were in existence when the earliest of the Pahrump strata were deposited and that they persisted through all of Pahrump time. Roberts (this volume) has shown, for example, that the Crystal Spring Formation displays (1) a southward fining in average grain size and in maximum size of clasts in the sandstone-conglomerate arkosic strata that compose its lower part, and (2) a general south-southwest orientation of cross-bedding in conglomeratic strata in that unit (Fig. 3D). The Beck Spring Dolomite, which in the central part of the trough is free of siliceous detrital beds, contains two tongues of fine-grained quartzite in exposures along the indicated site of the basin margin immediately south of Ashford Canyon on the west face of the Black Mountains.

The Kingston Peak Formation, which remains to be studied in detail, shows evidence of the Nopah upland by a marked northward thinning (Fig. 4A), as well as by its content of debris derived from the Crystal Spring Formation and Beck Spring Dolomite. The thinning is well displayed in exposures (1) in the Black Mountains north of the highway between Jubilee Pass and the Ashford Mill site (L. A. Wright and B. W. Troxel, 1974, unpub. data), (2) in the Alexander Hills at the south end of Nopah Range (Wright, 1973), and (3) in the vicinity of the Excelsior talc mine on the northeastern side of the Kingston Range (Wright, 1968). It is caused in part by erosion that preceded the deposition of the Noonday Dolomite, in part by the fact that some units that were deposited in the central part of the basin were not received by the topographically higher fault blocks along the margins of the basin, and in part by a stratigraphic thinning of individual units (Fig. 2).

The southward change in facies of the Noonday Dolomite,

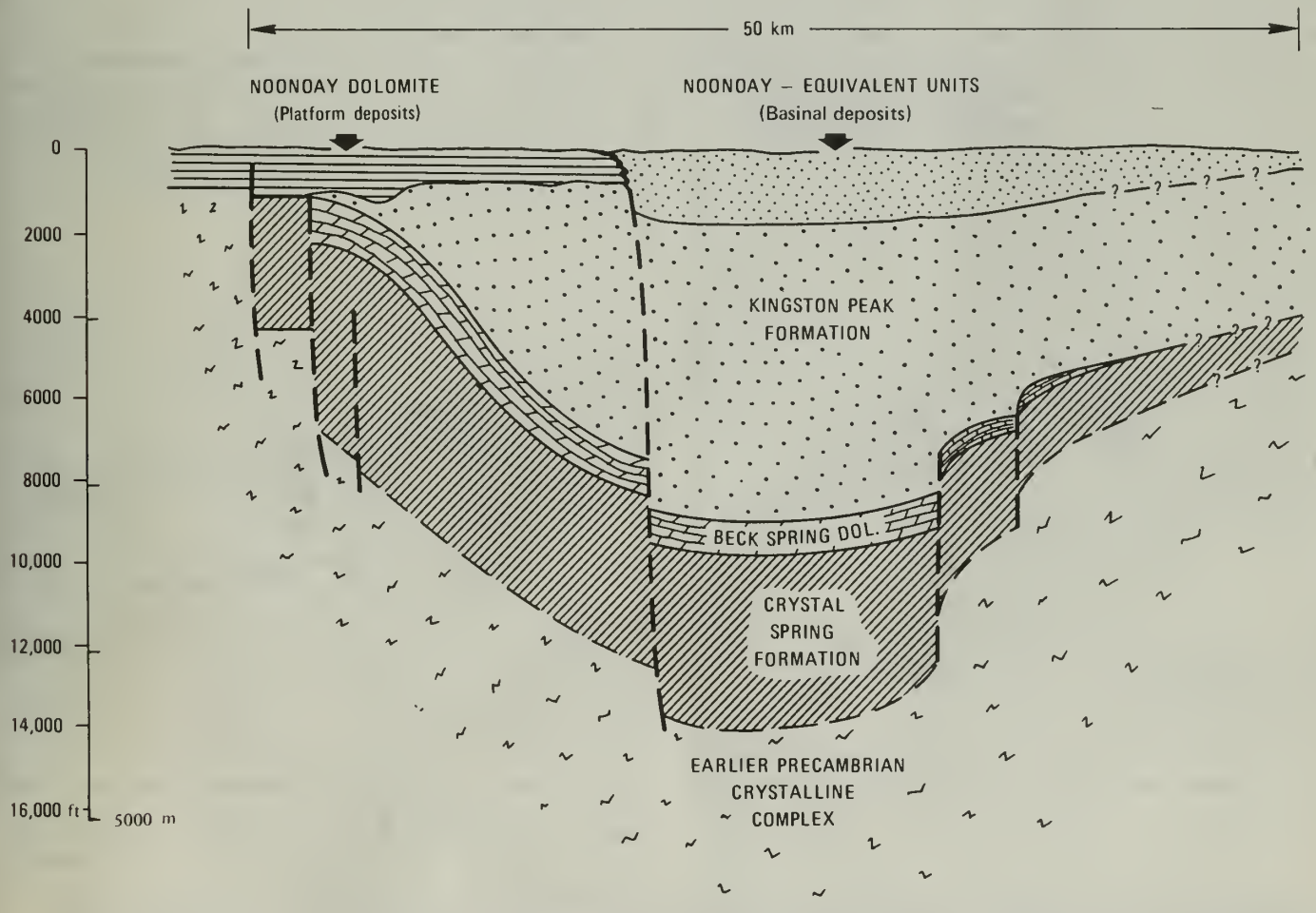

Figure 2. Diagrammatic cross section of Amargosa aulacogen at end of deposition of Noonday Dolomite and equivalent basinal sediments.

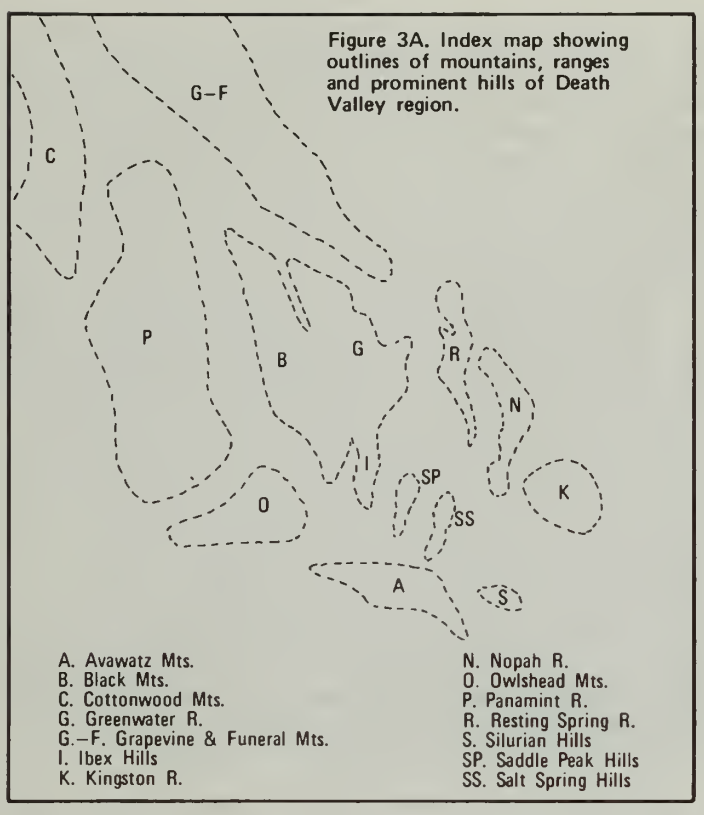

Figure 3A. Index map showing outlines of mountains, ranges and prominent hills of Death Valley region.

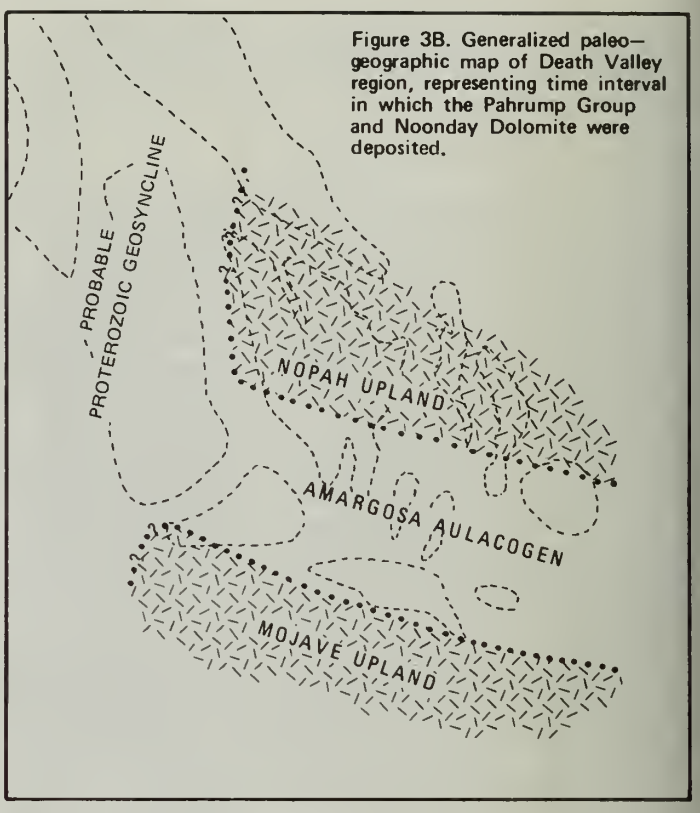

Figure 3B. Generalized paleogeographic map of Death Valley region, representing time interval in which the Pahrump Group and Noonday Dolomite were deposited.

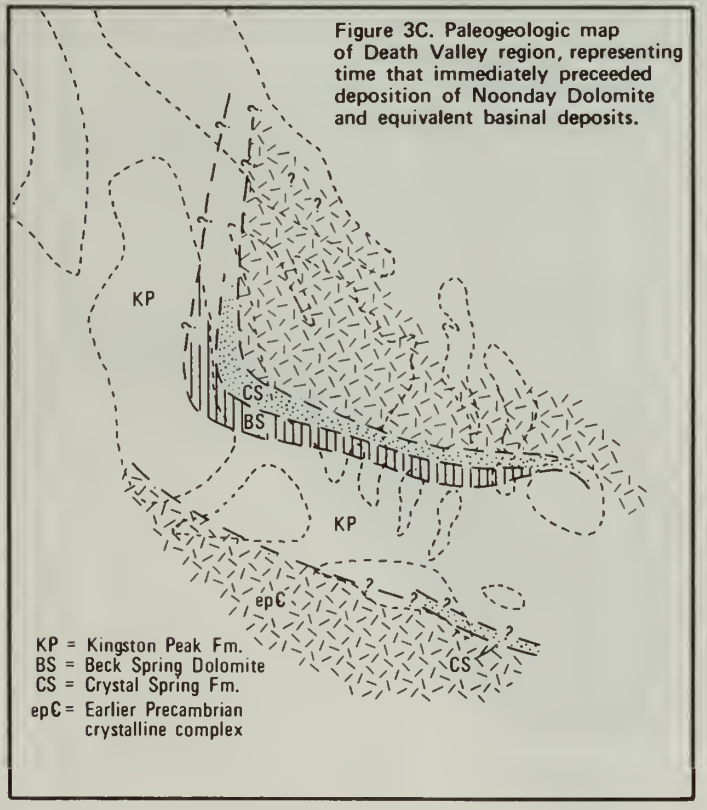

Figure 3C. Paleogeologic map of Death Valley region, representing time that immediately preceeded deposition of Noonday Dolomite and equivalent basinal deposits.

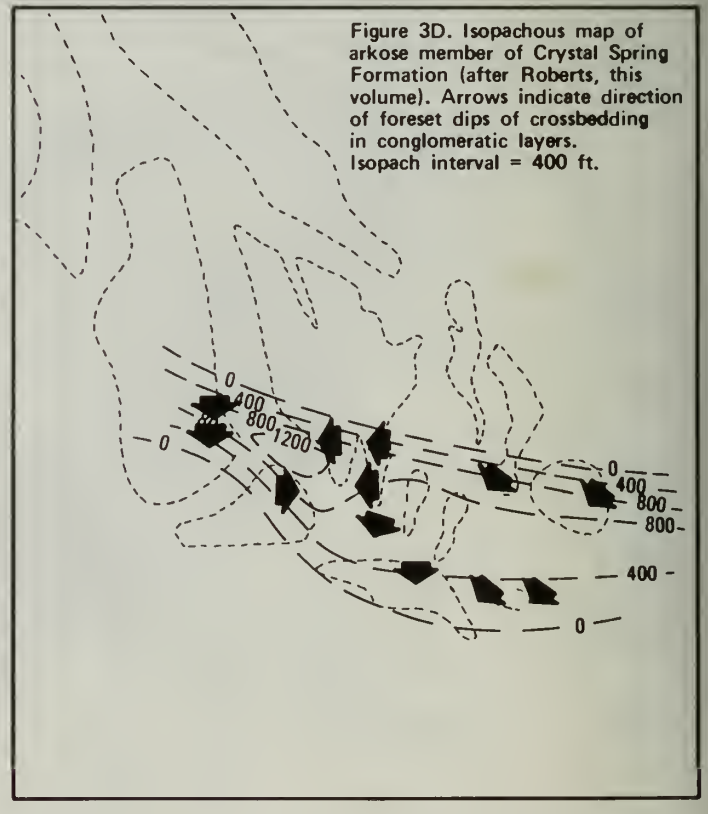

Figure 3D. Isopachous map of arkose member of Crystal Spring Formation (after Roberts, this volume). Arrows indicate direction of foreset dips of crossbedding in conglomeratic layers. Isopach interval = 400 ft.

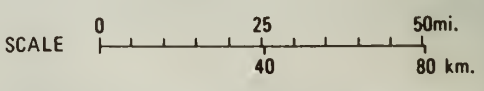

Figure 3. Paleogeographic maps of times when Pahramp Group and Noonday Dolomite were deposited, and isopachous map of Crystal Spring Formation.

from the platform carbonate to basinal strata (Williams and others, this volume), records a still later phase in the stratigraphic-tectonic development of the Amargosa aulacogen and its northern margin. The exposures of the Noonday that extend from the southwestern part of the Panamint Range for about 140 km south-southeastward (Figs. 2 and 4B) consist almost entirely of dolomite that has formed in the presence of algal mats. At various localities these platform deposits, which ordinarily are 200 m or less thick, can be traced abruptly into basin deposits two to three times at thick. In most places the latter are composed, in upward succession, of arkosic sandstone and siltstone in which graded bedding is common, thinly bedded clastic limestone and dolomite, and an almost structureless detrital unit of dolomite and quartz. Clastic dolomite and quartz also compose a strongly cross-bedded sandstone unit that overlies both the dolomite of the platform and the structureless unit of the basin and marks the transition from the Noonday Dolomite and equivalent units into the overlying Johnnie Formation.

A northerly to northeasterly source for the dolomite and quartz grains of the uppermost of the basinal units and the cross-bedded unit must be invoked to explain the intimate association of the two minerals, the quartz grains originating beyond the limits of the carbonate cover of the platform, passing over it, and mixing with clastic dolomite where the cover was being destroyed by wave or stream action. Measurements of current directions in the cross-bedded strata indicate a generally east-southeast strike for the paleoslope, essentially unchanged from the strike of the slope upon which the lowermost of the Pahrump strata were deposited. Indeed, measurements made to date of cross-bedding orientation in sandstone throughout the Johnnie Formation and Stirling Quartzite suggest that this strike persisted during the deposition of both (Fig. 4D). As these formations are exposed at various localities for many kilometers to the northeast of the basin, the source areas of the detrital material must have been much more distant from the basin site than the source areas of the detrital strata of the Pahrump units.

MOJAVE UPLAND

Evidence for exact location of the southern margin of the Amargosa aulacogen is less clear than that for the site of the northern margin, as Mesozoic and Cenozoic deformational and intrusive events have obscured the Precambrian record there. In the Silurian Hills, however, a conformable section of Pahrump strata, at least 2,220 m thick, has been identified (Kupfer, 1960; Wright and Troxel, 1966). The lower 300 m of this section is lithologically correlative with the lower and middle units of the Crystal Spring Formation as recognized elsewhere (Roberts, this volume). An apparently complete and thicker occurrence of the Crystal Spring also is exposed near the mouth of Sheep Creek Canyon on the north slope of the Avawatz Mountains. The distribution of a complex of post-Pahrump sedimentary units and Mesozoic(?) diorite, on either side of the valley that separates the Avawatz Mountains from the Silurian Hills, suggests right-lateral displacement of about 8 mi (13 km) and thus a strong possibility that these two Pahrump localities were once approximately opposite each other (Fig. 5D) on either side of the southern extension of Death Valley.

Exposures and intact sections of Pahrump- and Noonday-equivalent units in the southern part of the study area, although relatively few, indicate the existence of a southerly positive from which much detritus was fed into the Amargosa aulacogen (Fig. 2).

The earliest evidence of a southerly tectonic high is the southward thinning of the shallow-water arkosic lower part of the Crystal Spring (Roberts, this volume; Fig. 2), although Roberts has concluded that most or all of this detritus was derived from the northern upland. Quartzite layers in the middle part of the Crystal Spring Formation in the Silurian Hills (Kupfer, 1960) are interlayered with subordinate carbonate strata, the lower of which are dolomite and the upper of which are limestone. The clastic material is attributable to a southerly source. To the north, this part of the formation generally consists of two carbonate members, a lower member of siliceous dolomite and an upper limestone member containing stromatolites.

The higher part of the section in the Silurian Hills, as Kupfer (1960) observed, consists mostly of sandstone and conglomerate. Although he assigned the entire section to the Pahrump Group, he hesitated to make firm correlations with the established formations of the Pahrump. He noted, however, that a limestone unit, about 30 m thick and about 780 m above the base of the section, may be correlative with the Beck Spring Dolomite. He also observed that, in general, the section of the Silurian Hills contains coarser clastic rocks and a smaller proportion of carbonate material than those farther north. He thus proposed a southerly source for the clastic material.

Subsequent observations of the more northerly of the Pahrump occurrences support Kupfer's observations and interpretations to the extent that they apply to the upper part of the Crystal Spring Formation and the Beck Spring Dolomite. In the Ibex-Saratoga Hills the Beck Spring Dolomite shows a marked southward increase in insoluble material in the form of shaly and sandy layers. The upper units of the Crystal Spring in the central and northern parts of the basin are distinctly finer grained and contain a higher proportion of carbonate material than do the strata beneath the aforementioned limestone unit in the Silurian Hills.

Although most of the clastic material in the Kingston Peak Formation is traceable to the northerly source area, Troxel (1967) has recognized a southern Kingston Peak facies which contains clasts of metamorphic rocks unlike those in the rest of the formation and apparently derived from the south (Fig. 4A). Troxel has since observed that the two facies intertongue in a more complex fashion than his published description implies.

A southern source area also is indicated for the unit of arkosic sandstone and siltstone in the succession of strata that forms the basinal equivalent of the Noonday Dolomite (Fig. 4B). As observed by Williams and others (this volume), this unit contains bottom markings that evidence northerly transport, displays a northerly fining, and, in its northernmost exposures, shows an on-lapping relationship with the lower dolomite of the platform. The Nopah upland, at that time, was a low area largely or entirely covered by carbonate rock and thus could not have provided detritus of arkosic composition.

Stratigraphic-Tectonic Environments of the Amargosa Aulacogen

In brief, the various stratigraphic and structural features of the Amargosa aulacogen record a history divisible into two stages. Each is characterized by (1) an early period of erosion of the northern (Nopah) upland concurrent with clastic sedimentation in the subsiding trough, and (2) a later period of stability when that upland became a surface of low relief to serve as a platform and receive a cover of carbonate rocks. The southern (Mojave) upland apparently also was elevated with respect to sea level in two stages, one preceding and the other following the deposition of the Beck Spring–equivalent(?) unit of limestone in the Silurian Hills. These were approximately

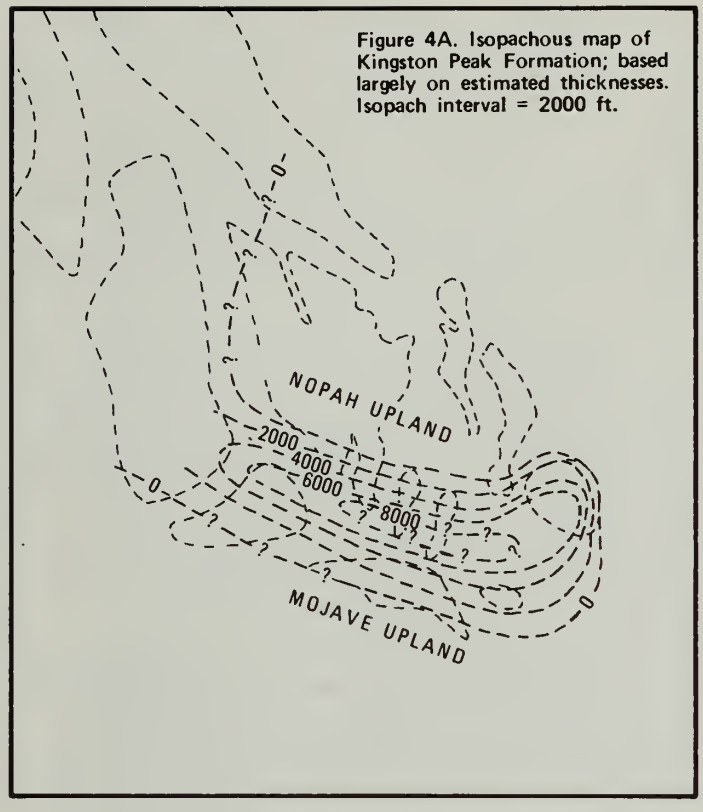

Figure 4A. Isopachous map of Kingston Peak Formation; based largely on estimated thicknesses. Isopach interval = 2000 ft.

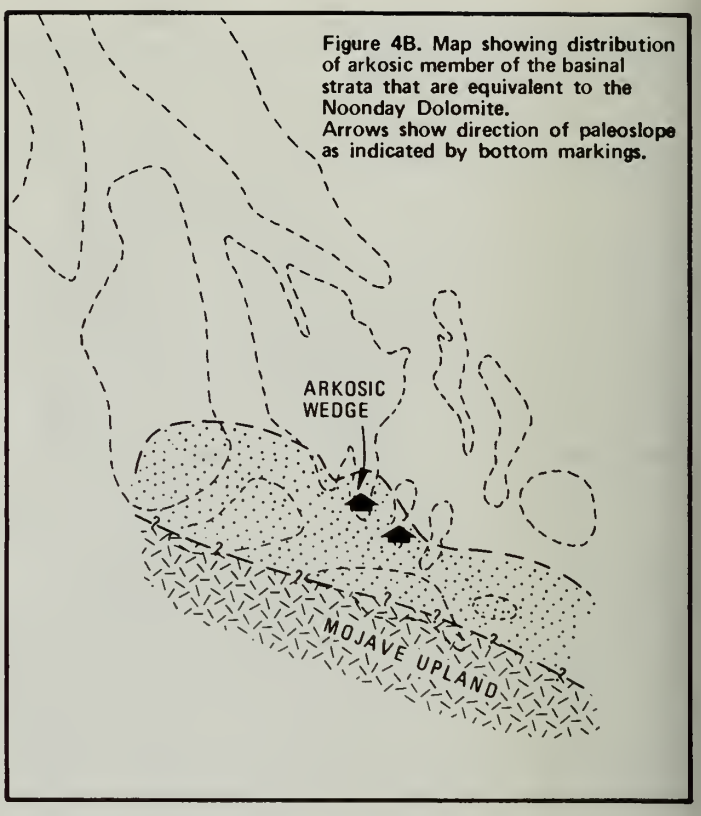

Figure 4B. Map showing distribution of arkosic member of the basinal strata that are equivalent to the Noonday Dolomite.
Arrows show direction of paleoslope as indicated by bottom markings.

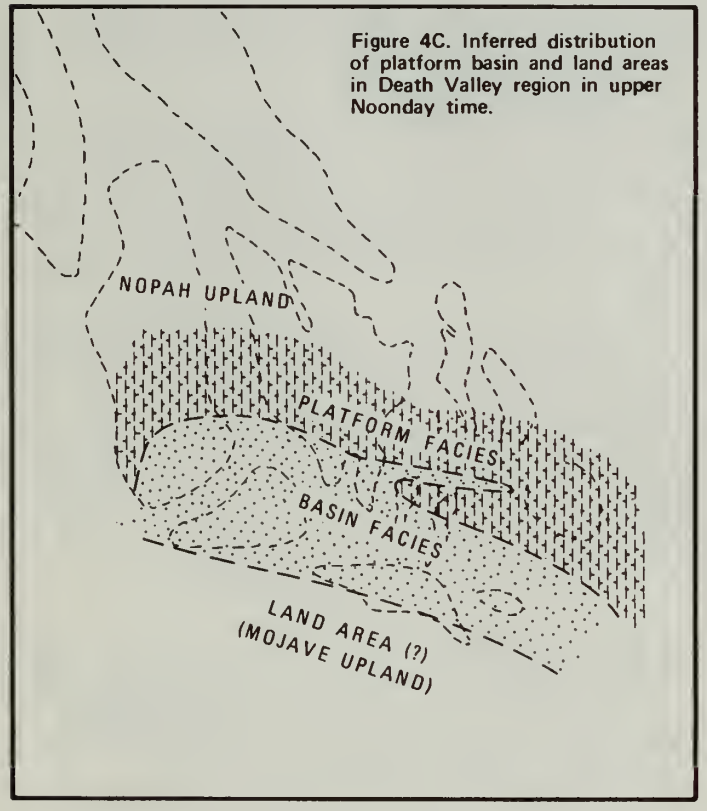

Figure 4C. Inferred distribution of platform basin and land areas in Death Valley region in upper Noonday time.

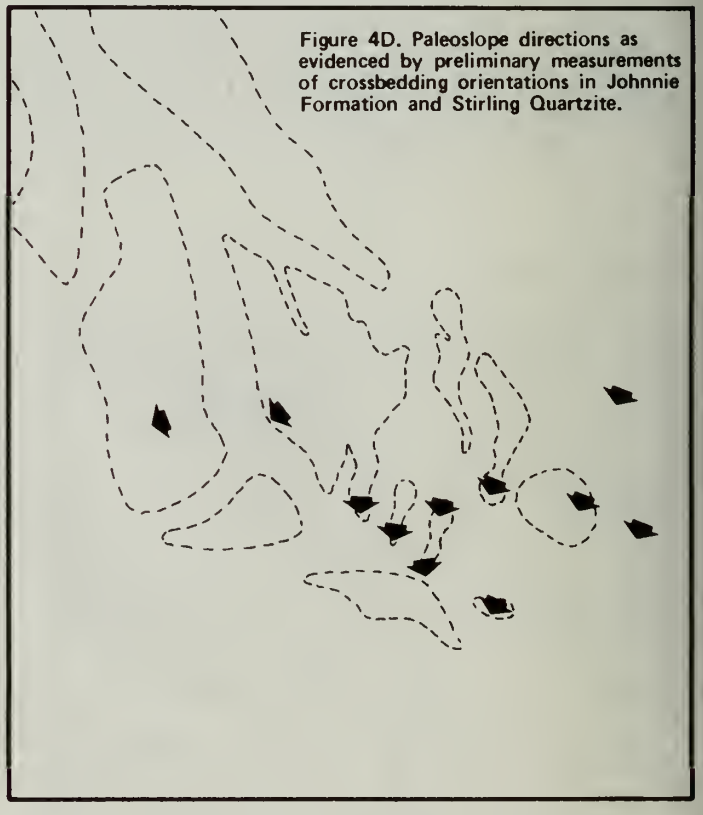

Figure 4D. Paleoslope directions as evidenced by preliminary measurements of crossbedding orientations in Johnnie Formation and Stirling Quartzite.

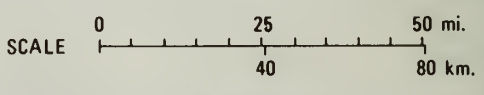

Figure 4. Isopachous map of Kingston Peak Formation, distribution map of Noonday Dolomite and equivalent basinal facies, and paleoslope directions of Johnnie and Stirling, Death Valley region. See Figure 3a for identification of mountains.

concurrent with the two erosional periods of the Nopah upland. But the trough received detritus from the northerly source considerably earlier than it did from the southerly source. The latter, however, may then have remained continually above sea level and, almost continually, a source area for clastic sediment. If it did, at times, receive carbonate platform deposits, a record of the covering remains unobserved either in situ or in the basinal detritus.

The two stages, although grossly similar, differed from each other in several major respects. The first stage, as recorded in the sediments of the Crystal Spring Formation and Beck Spring Dolomite, was featured by relatively slow subsidence compared with the subsidence rate during the second stage and shallow-water environments. Shallow water is indicated by abundant cross-bedding, ripple marks, and mud cracks in the units of clastic strata in the Crystal Spring, by algal stromatolites in the carbonate members of the Crystal Spring, and by algal stromatolites and layers of oolite in the Beck Spring. A progressive slowing of subsidence, with respect to sea level, and a concurrent lowering of the Nopah upland are indicated by the overall upward fining displayed by the clastic units, and by the deposition, in the later part of the first stage, of carbonate units across the trough site and onto the Nopah upland, which had been eroded to a platform. The emplacement of bodies of basic igneous rock, represented by the sill-forming diabase of the Crystal Spring Formation and diabase dikes in the crystalline complex, was confined almost entirely to the first stage, although basalt flows have been observed locally in the Kingston Peak Formation (A. L. Albee, personal communs.), as have thin layers of pale, fine-grained material that may be altered volcanic ash.

The second stage, which is recorded in the Kingston Peak Formation, the Noonday Dolomite, and the Noonday-equivalent basinal strata, on the other hand, was marked by higher marginal relief than is evidenced in the first stage, and by rapid subsidence of the trough. The central part received deep water sediment throughout the stage.

The Kingston Peak and Noonday-equivalent strata of the central trough consistently show evidence of deposition below wave base. Most display graded bedding, some are thinly and continuously laminated, others are massive and essentially structureless, and all are devoid of the shallow-water features that characterize the sedimentary rocks of the first stage. The conglomeratic bodies of the Kingston Peak Formation that lie along the trough margins have long been believed to represent alluvial fans (Hewett, 1956; Kupfer, 1960).

In contrast to the modest marginal uplift and basin sinking of the first stage, which diminished with time, the basinal sinking and marginal relief related to the first half of the second stage apparently increased, as evidenced in a general upward coarsening of the Kingston Peak clastic sediments and an increase in the proportion of graded beds of sand. Conglomerate layers high in that formation contain clasts as much as 0.3 km long. Boulders associated with the deep-water turbidite and diamictite layers of the Kingston Peak, some of which are striated, are believed by some to constitute evidence of glacial rafting, but their presence in deep-water sediments might also be explained by transportation in dense submarine debris flows.

In the later part of the second stage, the Nopah upland was again reduced to a surface of low relief, but it was featured along its southern margin by ridges, as much as 300 m high, of resistant conglomerate of the Kingston Peak Formation. On this platform was deposited the algal carbonate of the Noonday Dolomite, while the trough was receiving the deep-water basinal arkosic deposits derived from the Mojave upland. Still later, the basin was filled and the platform was largely covered by the dolomite-quartz sandstone and breccia transported southward.

That the Amargosa aulacogen is open and joins on the west to a geosyncline that was simultaneously receiving sediments like those of the aulacogen is suggested by the distribution of the most westerly and northerly exposures of units correlative with the Pahrump Group. These units occur in a belt that lies at a high angle to the aulacogen axis and extends along the west side of the Panamint Range as far north as Tucki Mountain and thence to the northern part of the Funeral Mountains on the northeast side of Death Valley (Fig. 3B). They remain to be examined in detail, but are known to comprise a lower part composed of siliceous clastic and carbonate rocks intruded by basic sills and an upper part composed of conglomerate with carbonate clasts. Farther to the west and north only younger rocks are exposed.

In summary, the various structural, stratigraphic, and igneous features that compose the Pahrump-Noonday terrane and its two-stage evolution correspond in general to the distinguishing features attributed by Hoffman and others (1974) to a typical aulacogen. The trough that received the basinal Pahrump and basinal Noonday-equivalent units was long-lived, being evidenced in stratigraphic units that cover the approximate time span of 1.4 to 0.7 b.y. If we have correctly interpreted the distribution of the most westerly and northerly of the Pahrump exposures, the trough joins with a geosyncline and lies at a high angle to the geosynclinal margin. The structural evolution of the trough and the nature of the sediments it received was controlled primarily by vertical movements on faults that bound the blocks on the trough margin, moderate in the earlier stage and strong during most of the later stage of trough development. Igneous activity, represented entirely by basic rock, occurred mostly in the earlier stage when dikes and sills of diabase were emplaced.

Aulacogen-Miogeosyncline Transition

The problem of identifying the record of transition from an aulacogenic to a miogeosynclinal environment in the Precambrian strata east of the Death Valley region remains unresolved, awaiting the acquisition of further data pertinent to the depositional environments of the Johnnie, Stirling, and Wood Canyon. When the quartzite member in the middle of the Johnnie (Fig. 1) was deposited, the platform-trough boundaries of the proposed aulacogen were apparently unexpressed topographically as the unit can be traced through the site of the trough and the immediately bordering uplands with no obvious variations except a thinning southward or southwestward. But the south to southwest paleoslope that is strongly suggested by paleocurrent directional features throughout the Johnnie is oriented similarly to the paleoslope upon which the earliest of the Pahrump strata were deposited. Measurements of limited numbers of directional features in the Stirling Quartzite suggest continuation of the south to southwest paleoslope for that formation in the Death Valley region, although the source or sources of the detrital quartz probably lay well beyond the sites of the aulacogen margins.

As Diehl (this volume) suggests, the principal indication of the change in provenance may well be the change to a northwesterly orientation in paleoslope evidenced in the cross-bedding and carbonate units thickening through most of the Wood Canyon Formation in the sections that he has studied (Fig. 5, A and B). As only the arkose and arkosic conglomerate beds low in the middle member of the Wood Canyon display cross-bedding that is consistently oriented southwestward, these possibly constitute the final record of uplift related to the pre-miogeosynclinal provenance.

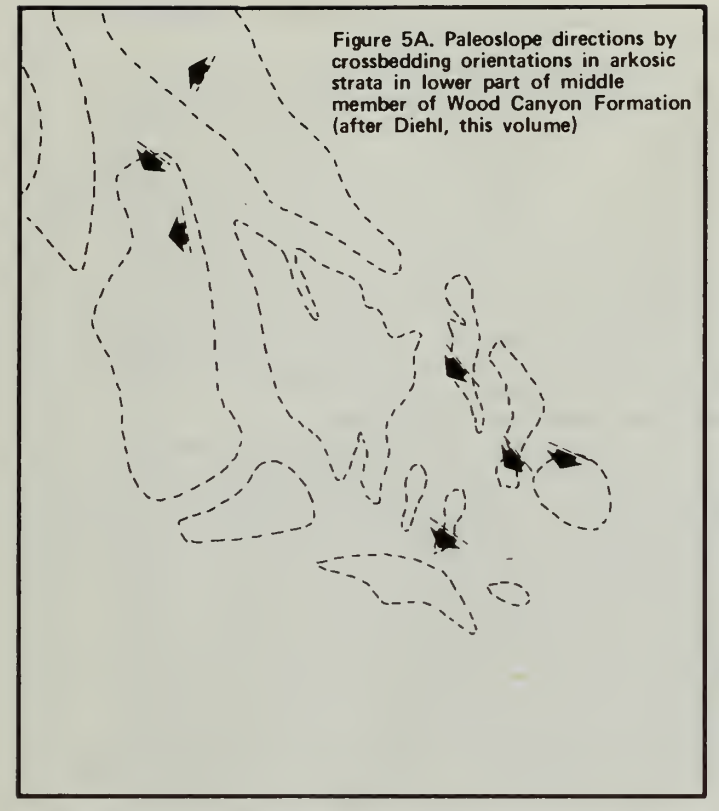

Figure 5A. Paleoslope directions by crossbedding orientations in arkosic strata in lower part of middle member of Wood Canyon Formation (after Diehl, this volume)

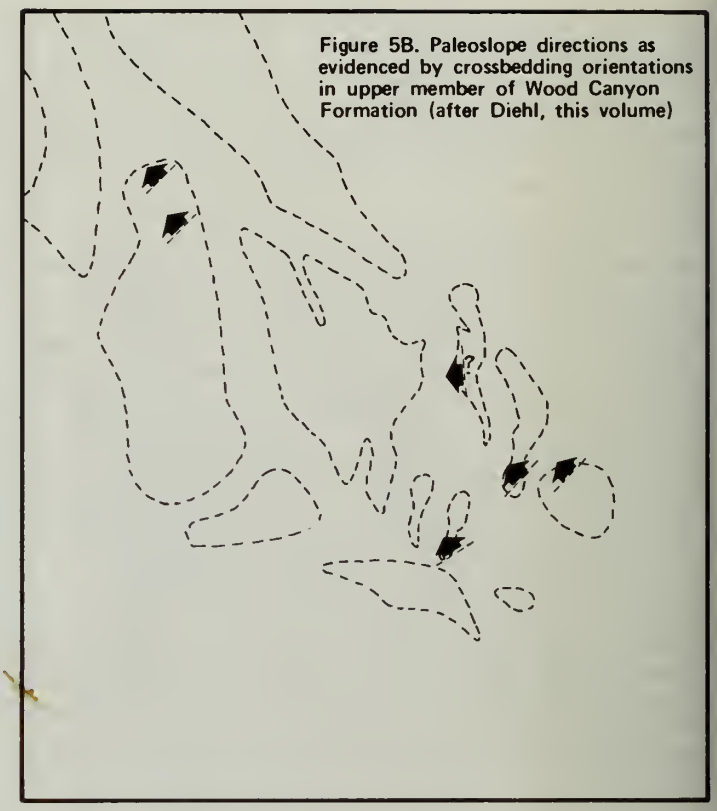

Figure 5B. Paleoslope directions as evidenced by crossbedding orientations in upper member of Wood Canyon Formation (after Diehl, this volume)

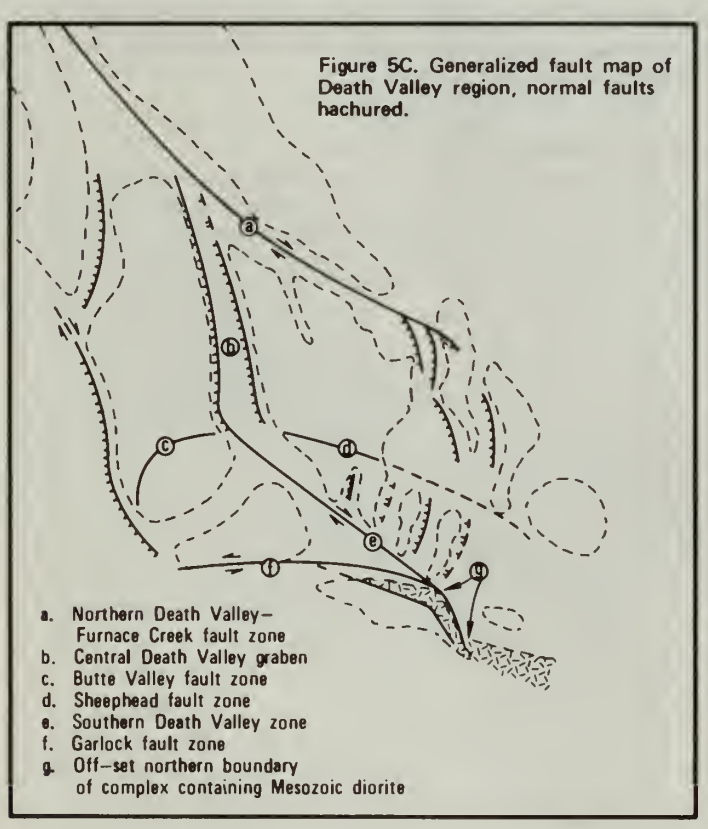

Figure 5C. Generalized fault map of Death Valley region, normal faults hachured.

a. Northern Death Valley–Furnace Creek fault zone
b. Central Death Valley graben
c. Butte Valley fault zone
d. Sheephead fault zone
e. Southern Death Valley zone
f. Garlock fault zone
g. Off–set northern boundary of complex containing Mesozoic diorite

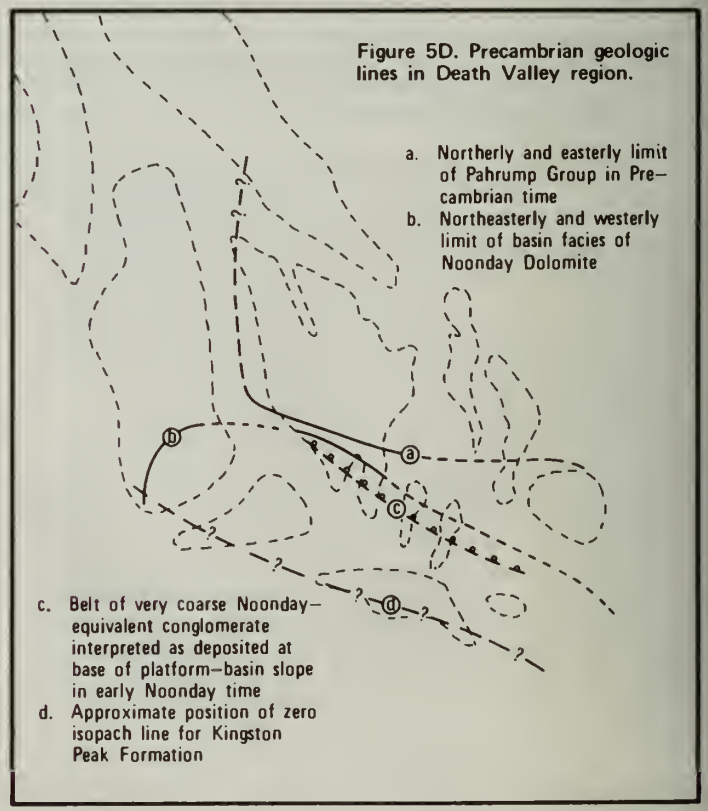

Figure 5D. Precambrian geologic lines in Death Valley region.

a. Northerly and easterly limit of Pahrump Group in Pre–cambrian time
b. Northeasterly and westerly limit of basin facies of Noonday Dolomite
c. Belt of very coarse Noonday–equivalent conglomerate interpreted as deposited at base of platform–basin slope in early Noonday time
d. Approximate position of zero isopach line for Kingston Peak Formation

SCALE 0 25 50 mi.
40 80 km.

Figure 5. Paleoslope directions of Wood Canyon Formation, fault map, and Precambrian geologic lines, Death Valley region. See Figure 3a for identification of mountains.

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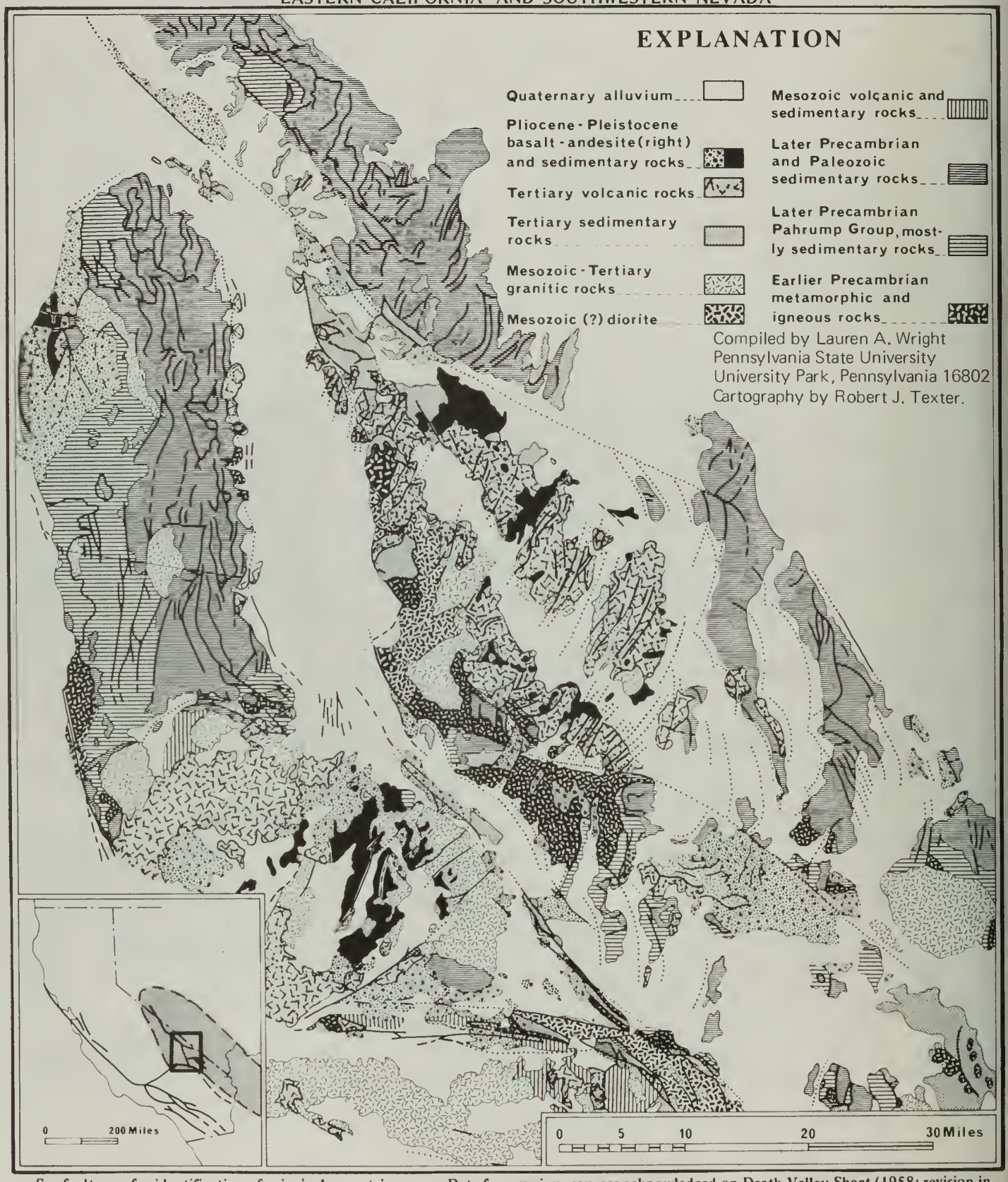
This research was supported by NSF Grant GA-16119.

High-altitude oblique aeriol photograph to south-southwest across the southern end of Deoth Volley. Avowatz Mountoins barely obove centrol part of photogroph. South end of Nopoh Ronge lies in lower center; Alexonder Hills (with thin white streok) lie slightly forther south. Amargoso River flows south from lower right of photogroph through potch of deformed pole Tertiory loke beds (just right of center foreground), then through deformed Tertiory-Quaternory grovel into the floor of southern Deoth Volley. The river then flows westword ot the north edge of the fon extending from the Avawotz Mountoins ond flows northword neor right edge of photogroph. The eost-trending Gorlock foult ond northwest-trending Southern Deoth Volley foult intersect along the north side of the Avawatz Mountoins.

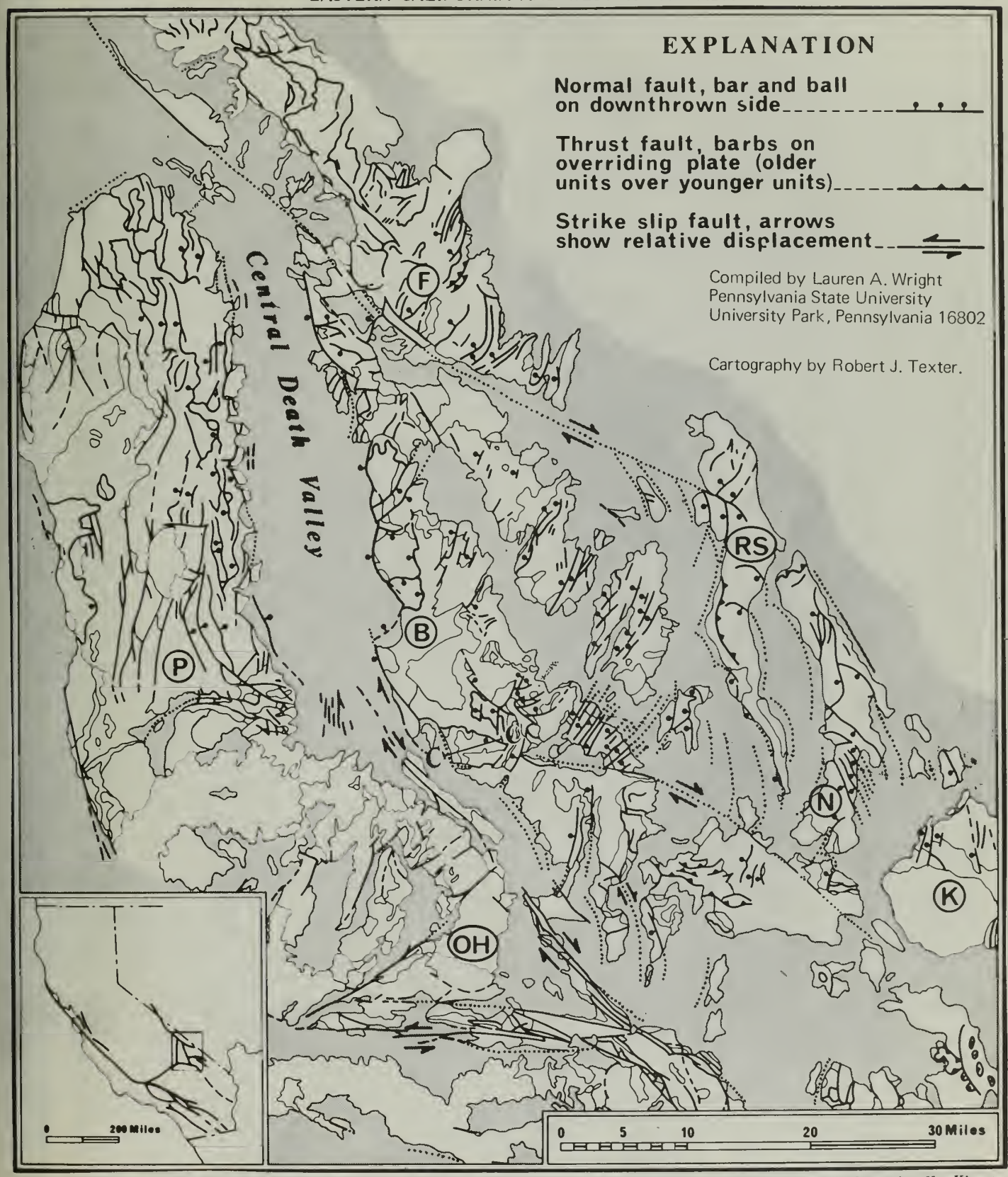
The Mojove River flows from the Son Bernordino Mountoins (dark ridge in upper center) ocross the Mojove Desert to Boker (on left side of dark hills beyond and left of Avowotz Mountoins). From Boker an oncient river channel continued northword to the dork hills (Solt Spring Hills) in center of photogroph. *Photo U.S. Air Force 374R 199, 6 September 1968; courtesy of the U.S. Geological Survey.*

GEOLOGIC MAP OF THE REGION OF CENTRAL AND SOUTHERN DEATH VALLEY, EASTERN CALIFORNIA AND SOUTHWESTERN NEVADA


EXPLANATION
Quaternary alluvium
Pliocene-Pleistocene basalt-andesite(right) and sedimentary rocks
Tertiary volcanic rocks
Tertiary sedimentary rocks
Mesozoic-Tertiary granitic rocks
Mesozoic (?) diorite
Mesozoic volcanic and sedimentary rocks
Later Precambrian and Paleozoic sedimentary rocks
Later Precambrian Pahrump Group, mostly sedimentary rocks
Earlier Precambrian metamorphic and igneous rocks
Compiled by Lauren A. Wright
Pennsylvania State University
University Park, Pennsylvania 16802
Cartography by Robert J. Texter.
0 200 Miles
0 5 10 20 30 Miles

See fault map for identification of principal mountain ranges. Data from various sources acknowledged on Death Valley Sheet (1958; revision in preparation, 1973), and Trona Sheet, Geologic Map of California, California Division of Mines and Geology. Reproduced by permission and modified from *Gravity and Tectonics*, edited by Kees A. De Jong and Robert Scholtn (New York: John Wiley & Sons), 1973.

FAULT MAP OF THE REGION OF CENTRAL AND SOUTHERN DEATH VALLEY, EASTERN CALIFORNIA AND WESTERN NEVADA



Shaded pattern denotes Quaternary alluvium; blank areas denote pre-Quaternary rocks. B = Black Mountains; F = Funeral Mountains; K = Kingston Range; N = Nopah Range; OH = Owlshead Mountains; P = Panamint Range; RS = Resting Spring Range. Reproduced by permission and modified from *Gravity and Tectonics*, edited by Kees A. De Jong and Robert Scholtn (New York: John Wiley & Sons), 1973.

Low altitude oblique aerial photograph across Death Valley northeast of Gropevine Mountains between mouth of Titus Canyon (left) and Titanothere Canyon (right). Pale rocks just below skyline are Tertiary rocks, mostly volcanic. Stratified rocks in Gropevine Mountains range in age from early Cambrian (right) to late Paleozoic (left) and are highly deformed. See article that follows. *Photo W16 by John H. Maxson; courtesy of the National Park Service.*

Geology of the Grapevine Mountains, Death Valley, California: A Summary

Mitchell W. Reynolds[1]

INTRODUCTION

The Grapevine Mountains, limited on the south by Boundary Canyon and on the north by Grapevine Canyon, form the eastern wall of northern Death Valley (Fig. 1). The mountains are rugged and have poor accessibility, but their high relief and barrenness provide excellent exposures of the complicated geology characteristic of the Death Valley region. Upper Precambrian and Paleozoic rocks exposed in the core of the mountains are the most northerly complete section of miogeosynclinal facies strata on the east side of the northern Death Valley–Furnace Creek fault zone. In the core of the mountains, a thrust fault of early or medial Mesozoic age juxtaposes upper Precambrian, Cambrian, and Ordovician rocks over middle and upper Paleozoic strata. Cenozoic rocks flank the core and include both the oldest dated Tertiary strata in the southwestern part of the Great Basin and an extensive succession of younger sedimentary and volcanic rocks. These and older strata are broken by late Cenozoic high-angle normal faults that flatten with depth east and west off the core of the range. In the southern part of the mountain core, folding and faulting of late Tertiary age produced a north-trending S-shaped fold system composed of a recumbent syncline and anticline. A low-angle fault, along which rocks are rotated and extended, replaces the upright limb of the anticline, and the entire fold system is separated from underlying metamorphosed Precambrian rocks by a low-angle fault that also displaces Tertiary strata at shallow depths. Strike-slip faults of late Cenozoic age displace rocks and structures in the range, as well as Quaternary deposits in adjacent Death Valley.

In this summary of the geology of the Grapevine Mountains, emphasis is on the southern part of the mountains, since this area is accessible by road through Boundary and Titus Canyons. The geology of the Grapevine Mountains has been mapped largely by me; mapping by Cornwall and Kleinhampl (1964) in the southeasternmost part of the mountains and by L. A. Wright and B. W. Troxel (1971, written commun.) along the southern edge has provided a base for more detailed mapping and structural interpretations of those parts by me.

STRATIGRAPHY

Nearly 9,100 m (30,000 ft) of pre-Mesozoic and Cenozoic sedimentary and volcanic rocks are exposed in the Grapevine Mountains (Fig. 2). For about 9.6 km (6 mi) southeast of Boundary Canyon to Chloride Cliff, the oldest rocks crop out in the anticlinal core of the Funeral Mountains (Wright and Troxel, 1970; 1971, written commun.). These rocks belong to the Pahrump Group of Precambrian age. Northwestward, a nearly continuous succession of pre-Mesozoic rocks is preserved down structural plunge from Chloride Cliff to the north end of the Grapevine Mountains, where strata of Pennsylvanian age crop out. The lowest quarter of the succession, widely exposed in Boundary Canyon, is a thick sequence of terrigenous clastic rocks, whereas much of the remainder of the section is composed of carbonate rock with only thin, interbedded terrigenous units. Some of the Precambrian, Cambrian, and Mississippian rocks may have accumulated in moderately deep marine waters, but the carbonate and terrigenous rocks were generally deposited in shallow-marine and intertidal environments on a slowly subsiding shelf. Siltstone, sandstone, and conglomerate beds among upper Paleozoic rocks were derived from uplands presumably to the northwest and north resulting from the Antler orogeny.

Terrigenous strata in the pre-Mesozoic sequence have generally been the loci of failure during deformation. Preferred units of failure are designated by asterisks in Figure 2. The Stirling and Zabriskie Quartzites, for example, controlled positions of faulting during both Mesozoic and Cenozoic deformation: these formations form the sole of the allochthon of the major thrust fault of Mesozoic age in the northern part of the Grapevine Mountains, and form the soles of flat faults of Cenozoic age in Boundary Canyon. Recumbent folding, probably late Tertiary in age, in the southern Grapevine Mountains involved rocks only as old as the Stirling Quartzite, which acted as a sole for detachment above metamorphosed and deformed older rocks. High-angle faults of Cenozoic age are commonly flattened with depth to follow siltstone and quartzite beds in Paleozoic strata.

Cenozoic rocks crop out widely along the east flank and the northwest front of the Grapevine Mountains and locally near Death Valley Buttes, west of the mouth of Boundary Canyon. Rocks on the northwest front differ in origin and age from those on the east and south. At the latter localities, strata of fluvial and lacustrine origin belonging to the Titus Canyon Formation of Oligocene and Miocene(?) age (Reynolds, 1969, 1974) rest unconformably on Paleozoic rocks and are, in turn, overlain unconformably by ash-flow tuff, lava flows, and sedimentary rocks of lacustrine and fluvial origin. This succession of volcanic and sedimentary rocks is about 22 to 20 m.y. old, and sources for most of the volcanic rocks lay within the Nevada Test Site east and northeast of the Grapevine Mountains.

Cenozoic rocks are exposed along the mountain front in the northern part of the Grapevine Mountains northwest of Fall Canyon and at the southern end of the mountains in the Kit Fox Hills (Fig. 1). Rocks northwest of Fall Canyon are divided into three parts—the lowest exposed part and the upper part are dominantly of fluvial origin, and rocks of the middle part are of lacustrine origin. The lower fluvial and lacustrine rocks are conformable, but the upper fluvial sequence lies unconformably on the lacustrine rocks. Thin, coarse basaltic intrusions occur in the lowest two units. Clasts in the lower fluvial sequence were derived from granitic intrusive rocks exposed at the north end and west of Death Valley; clasts of welded ash-flow tuff were eroded from units northeast and east in the Grapevine Mountains and beyond, but clasts of locally derived pre-Mesozoic rocks are not abundant. By contrast, the upper fluvial rocks are composed mainly of fragments eroded from pre-Mesozoic rocks in the Grapevine Mountains, onto which the fluvial strata lap or are faulted. The lower two units are here correlated with the Furnace Creek Formation as defined by McAllister (1970) and the upper unit with the Funeral Formation, both in the central part of Death Valley. Cenozoic lacustrine and fluvial rocks of the Kit Fox Hills west of Death Valley Buttes (Fig. 1) contain clasts derived from ash-flow tuff units of Miocene and early Pliocene age exposed east of Death Valley (Fig. 2), and hence are considered to be

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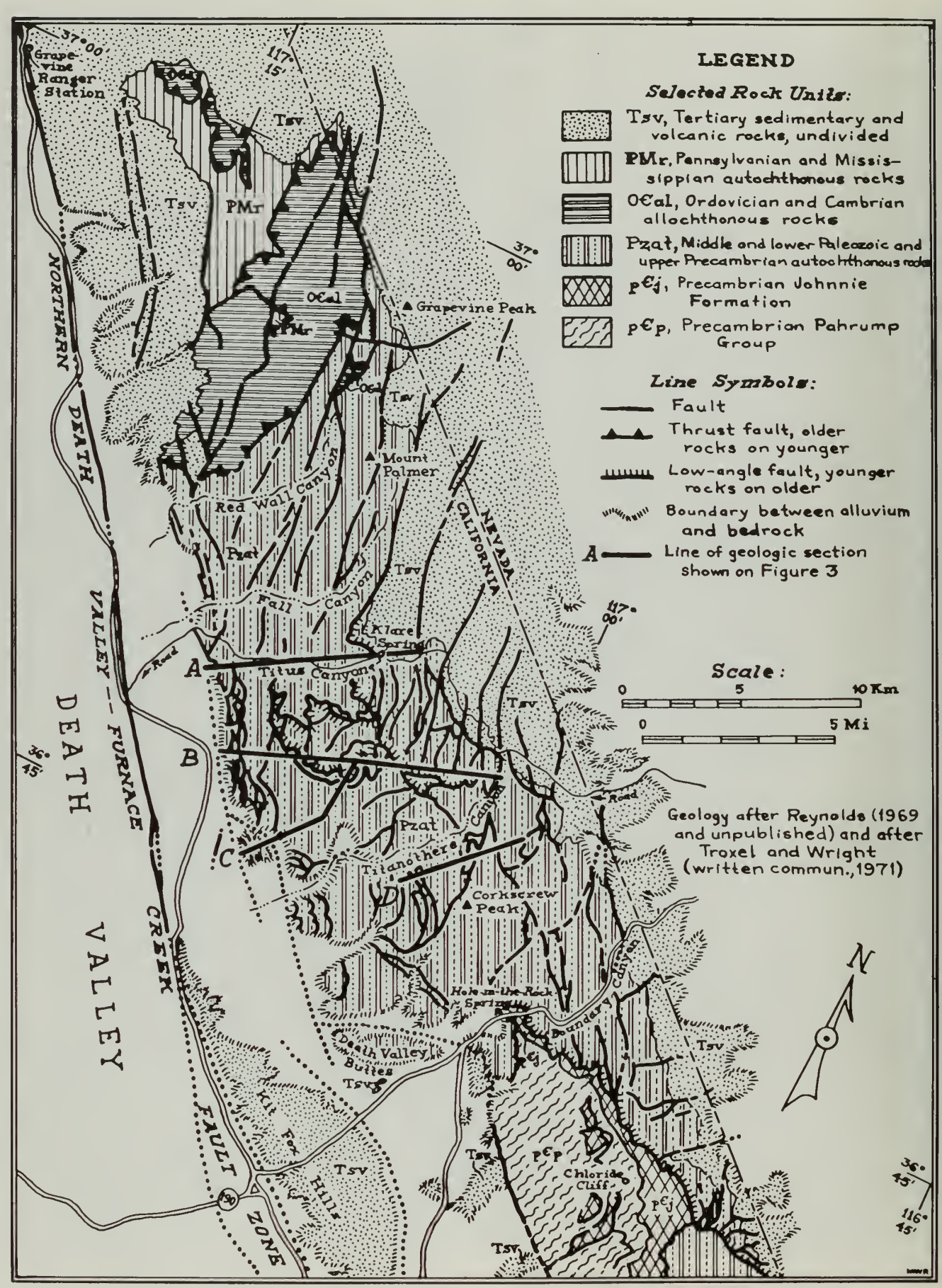


Figure 1. Generalized geologic map of the Grapevine and northern part of the Funeral Mountains, Death Valley, California, showing localities referred to in the text.

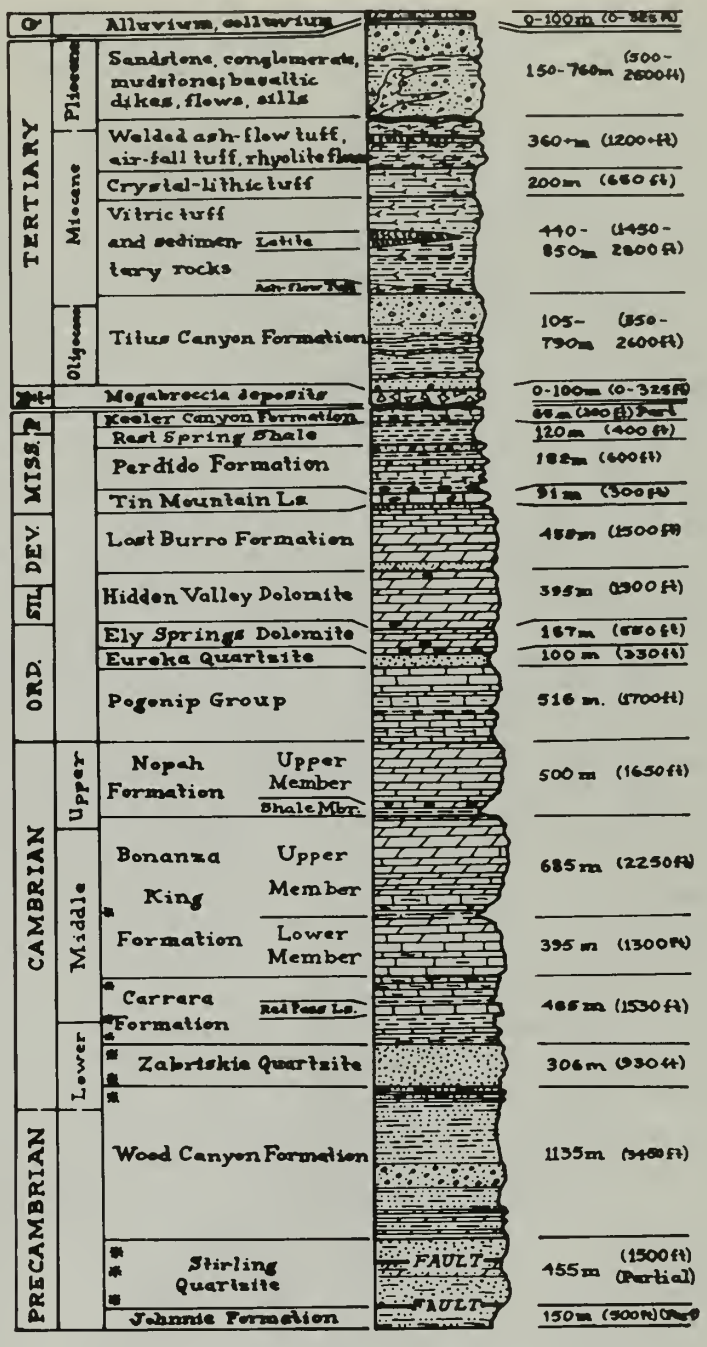
of middle or late Pliocene age, equivalent to the Furnace Creek Formation (compare with Hunt and Mabey, 1966, p. A57).

Alluvium eroded from the mountains spreads into Death Valley from major canyons and rills. Hunt and Mabey (1966) and Reynolds (1969) recognized four ages of alluvium northward from Boundary Canyon. The alluvium is both faulted against the mountain front from Titanothere to Red Wall Canyon (Fig. 1) and banked depositionally against several fault scarps, demonstrating intermittent movement along different segments of the mountain front. Alluvium adjacent to the mountains is unfaulted south of Titanothere Canyon to beyond Boundary Canyon, although locally warped. Clay, silt, and evaporite minerals are accumulating on a playa adjacent to the steep front of the Grapevine Mountains between Titus and Titanothere Canyons. The eastward displacement of the playa with respect to the axis of Death Valley toward the frontal scarp of the mountains suggests that that segment of the valley is tilting downward to the east. Both alluvium and playa sediments are displaced by faults along the northern Death Valley–Furnace Creek fault zone.

The general geologic structure of the Funeral and Grapevine Mountains is a broad anticlinorium with the deepest structural level exposed in the culmination near Chloride Cliff in the northern part of the Funeral Mountains. Progressively shallower structural levels are exposed northward across the Grapevine Mountains and southeastward across the southern part of the Funeral Mountains. Shallower structural levels are also exposed down the dip of the east flanks of the mountains and locally on the west flanks. The exposed structures developed during two principal times of deformation: the older during early Mesozoic time (probably Late Triassic or Early Jurassic [Burchfiel and others, 1970]), and the younger encompassing middle and late Cenozoic time. Different styles of deformation characterized by thrust faulting in which upper Precambrian and lower Paleozoic rocks were thrust over middle and upper Paleozoic rocks. The later deformation, however, is primarily distinguished by normal faulting and doming associated with regional tectonic extension and igneous activity. The following summary describes structures of early Mesozoic age and emphasizes structures of Cenozoic age.

Structures of Mesozoic Age. Across the northern part of the Grapevine Mountains, rocks of Precambrian, Cambrian, and Ordovician age are thrust over rocks ranging in age from Ordovician into Pennsylvanian. The fault is called the Grapevine thrust fault (Reynolds, 1971). At the base of the allochthonous plate about 8 km (5 mi) west-northwest of Mount Palmer, strongly contorted beds of the Stirling Quartzite are tectonically interleaved with deformed beds of the Wood Canyon Formation. There these rocks rest on the autochthonous Perdido Formation and Rest Spring Shale of Mississippian and Pennsylvanian(?) age. North and west-southwest of Mount Palmer, the Wood Canyon Formation is the lowest exposed unit of the allochthon. Rocks of the upper plate generally dip north across the north end of the Grapevine Mountains, so that the Cambrian Zabriskie Quartzite, Carrara, Bonanza King, and Nopah Formations, and the Ordovician Pogonip Group successively form the sole of the upper plate. Rocks of the upper plate are folded in a tight north-trending anticline 3.2 km (2 mi) southwest of Grapevine Peak, in the core of which the Wood Canyon Formation is exposed, but otherwise the plate was not strongly deformed during thrusting.

The authochthon includes the entire succession of Paleozoic and Precambrian rocks exposed from outcrops of the Grapevine thrust fault southward up the regional plunge to the major culmination of the anticlinorium near Chloride Cliff. Immediately beneath the fault, middle and upper Paleozoic rocks are strongly deformed: 9.7 km (6 mi) west of Grapevine



*Denotes interval of common failure during folding and faulting. Thicknesses are approximate. MzT denotes Mesozoic(?) and Tertiary age. Data of Mitchell W. Reynolds.

Figure 2. Generalized columnar section of rocks exposed in the Grapevine Mountains, Death Valley, California.

Peak, the Mississippian Perdido Formation, the Rest Spring Shale, and the Pennsylvanian Keeler Canyon Formation are locally overturned west or northwest beneath the fault. This overturning suggests that the allochthon moved southeast and east. West of Mount Palmer the Devonian Lost Burro Formation is locally contorted in disharmonic folds beneath the thrust fault. Farther south the autochthon seems to have been little deformed during thrusting.

Stewart and others (1966) suggested that the Grapevine thrust fault may be equivalent to the Last Chance thrust fault that they described from extensive exposures west of Death

Valley. The Last Chance thrust fault is structurally the highest and youngest thrust fault and seems to have the greatest displacement in a succession of thrust faults west of Death Valley, including the Lemoigne, Gap Hills, Big Horn, Quartz Springs, and Racetrack thrust faults (McAllister, 1952; Hall and Stephens, 1962). I (1971) tentatively accepted the correlation of the Last Chance and Grapevine thrust faults and noted that the Grapevine thrust fault continues east from the Grapevine Mountains through exposures in the Bullfrog Hills to Bare Mountain, where Cornwall and Kleinhampl (1961) described a thrust fault at Meiklejohn Peak, which shows similar structural and stratigraphic relations. Although the specific surface of thrusting is not likely the same across the region, the exposed fault surfaces are probably part of the Last Chance thrust system. Thrust faults equivalent to the older described faults on the west side of Death Valley do not seem to be present on the east side in either the Grapevine Mountains or the northern part of the Funeral Mountains. Plutons of quartz monzonite that intrude the thrust faults west of Death Valley have approximate radiometric ages of 156 and 165 m.y., and Burchfiel and others (1970) consider that thrusting occurred during latest Paleozoic and early Mesozoic time, most likely during Middle Triassic to Early Jurassic time. That age of thrusting is also accepted for the Grapevine thrust fault.

No stratigraphic record of tectonic events younger than thrusting but older than Oligocene age is preserved in the Grapevine Mountains. Deformed Precambrian and Paleozoic rocks were tilted northward prior to erosion of the surface on which rocks of the Titus Canyon Formation (Oligocene and Miocene(?)) accumulated. That formation rests on Precambrian–Cambrian Wood Canyon Formation east of Chloride Cliff (L. A. Wright and B. W. Troxel, 1971, written commun.) and on progressively younger strata to the north where, near Grapevine Peak, it rests unconformably on allochthonous rocks of the Grapevine thrust fault. The relation suggests that the area of the Funeral Mountains was a structural high prior to the erosion and that the present anticlinorium represents a middle and late Cenozoic rejuvenation of that earlier structure.

Structures of Cenozoic Age. The most conspicuous geologic structures in the Grapevine Mountains and northern Funeral Mountains developed during Cenozoic time. Deformation was episodic in pulses between 20 and 16, ~14 and 13, and 11 and 7 m.y., and has been nearly continuous since (Reynolds, 1974). The first two episodes were characterized by uplift in the southern part of the area and by faulting along northeast and north trends. The episode between 11 and 7 m. y. seems to have been particularly intense, for, during that time, doming, recumbent folding, and extensive faulting along a north trend occurred. Intermittent younger deformation has produced much of the relief evident along the mountain fronts and right-lateral faulting that displaces Quaternary units in northern Death Valley.

In section view from north-northwest to south-southeast along the Grapevine Mountains and northern Funeral Mountains, the general structure that developed during Cenozoic time is an anticlinorium with a major culmination in the vicinity of Chloride Cliff and a second culmination between Mount Palmer and Grapevine Peak (Fig. 1). From east to west the ranges are anticlinal with tilted, faulted Tertiary rocks on the east and west flanks. Minimum structural relief between the culmination at Chloride Cliff and the northern end of the Grapevine Mountains is 10.6 km (6.6 mi). Southeast down plunge from Chloride Cliff the apparent structural relief exceeds 6 km (3.7 mi). Structural relief between the crest of the second culmination near Mount Palmer and the equivalent stratigraphic horizon in the adjacent depression that lies between Titus and Titanothere Canyons is about 3 km (1.9 mi). Part of the relief had developed prior to Oligocene time, but most developed in Pliocene time, as demonstrated by strata of early Pliocene age tilted to angles of 45° on the flanks of the folds.

About 8 km (5 mi) east-southeast of Chloride Cliff, high-angle faults that displace rocks as young as Pliocene flatten westward and northward with depth, merging to form a single low-angle fault (L. A. Wright and B. W. Troxel, 1971, written commun.). Progressively north along the east flank of the range, north-trending faults also flatten with depth to join the master low-angle fault. In its southernmost exposures that fault is within the Stirling Quartzite, but from about 5 km (3 mi) east of Chloride Cliff to Boundary Canyon the fault forms a surface of décollement between the Precambrian Johnnie Formation below and the Stirling Quartzite and younger strata above. Across that area, Tertiary, Cambrian, and Precambrian strata, in blocks bounded by the high-angle faults that flatten to join the master fault, are rotated downward to the east or northeast to angles as steep as 75°. The low-angle master fault is well exposed between the Stirling Quartzite and Johnnie Formation on the north side of Boundary Canyon from south of Hole-in-the-Rock Spring east for about 1 km (0.7 mi).

From Boundary Canyon north as far as Titus Canyon in the Grapevine Mountains, the Stirling Quartzite, Wood Canyon Formation, and Cambrian beds are folded in an S-shaped fold system that consists of a recumbent syncline on the west and a structurally higher upright-to-recumbent anticline on the east (Fig. 3). The low-angle fault visible beneath the Stirling in Boundary Canyon is the surface of décollement between the recumbently folded strata and metamorphosed deformed older rocks. A limestone unit in the lower part of the Carrara Formation outlines the tight core of the recumbent syncline on Corkscrew Peak, and the Stirling and Wood Canyon beds east for 2 km (1.3 mi) from Hold-in-the-Rock Spring outline the broader, lower part of the syncline. The anticline is defined by outwardly opposing dips in the Wood Canyon Formation exposed on the northwest side of the highway through Boundary Canyon about 6.5 km (4 mi) above the spring. In Boundary Canyon, the anticline is faulted downward approximately 1.6 km (1 mi) against the overturned east limb of the syncline.

Traces of the fold axes trend about N. 60° W. toward a salient in the S-shaped fold system near Titus Canyon; north of the salient they trend north. The folds plunge northwest off the culmination of the Funeral Mountains anticlinorium. The trace of the axial surface is exposed in progressively younger beds northwest from Boundary Canyon in the recumbent syncline, so that between Titanothere and Titus Canyons the Upper Cambrian Nopah Formation forms the core of the fold (B and C in Fig. 3). Beds of the Cambrian Bonanza King Formation are the youngest exposed on the upright limb of the anticline near Titus Canyon, but farther north progressively younger Paleozoic rocks are continuous above the Bonanza King. Otherwise, only older strata are exposed in the anticline, and generally the overturned limb on the west has been faulted out. The axial surface of the syncline dips 20° to 60° to the east; for about 5 km (3 mi) southeast from the mouth of Titus Canyon, the recumbent syncline has been rotated downward toward Death Valley so that the axial surface dips about 15° to the west (A and B in Fig. 3). At Boundary Canyon the axial surface of the anticline is nearly vertical, but it dips northeast at progressively lower angles toward Titanothere Canyon. The recumbent part of the anticline is fully preserved only north of Titus Canyon where the axial surface dips from 5° to 10° to the east (A in Fig. 3).

Between Titus and Titanothere Canyons, where the west-directed salient of the fold system coincides with the structural depression in the range, beds of the upright limb of the anticline are broken by numerous normal faults (B in Fig. 3). Along these faults the strata have been rotated downward to the west and effectively extended over a wider area. The

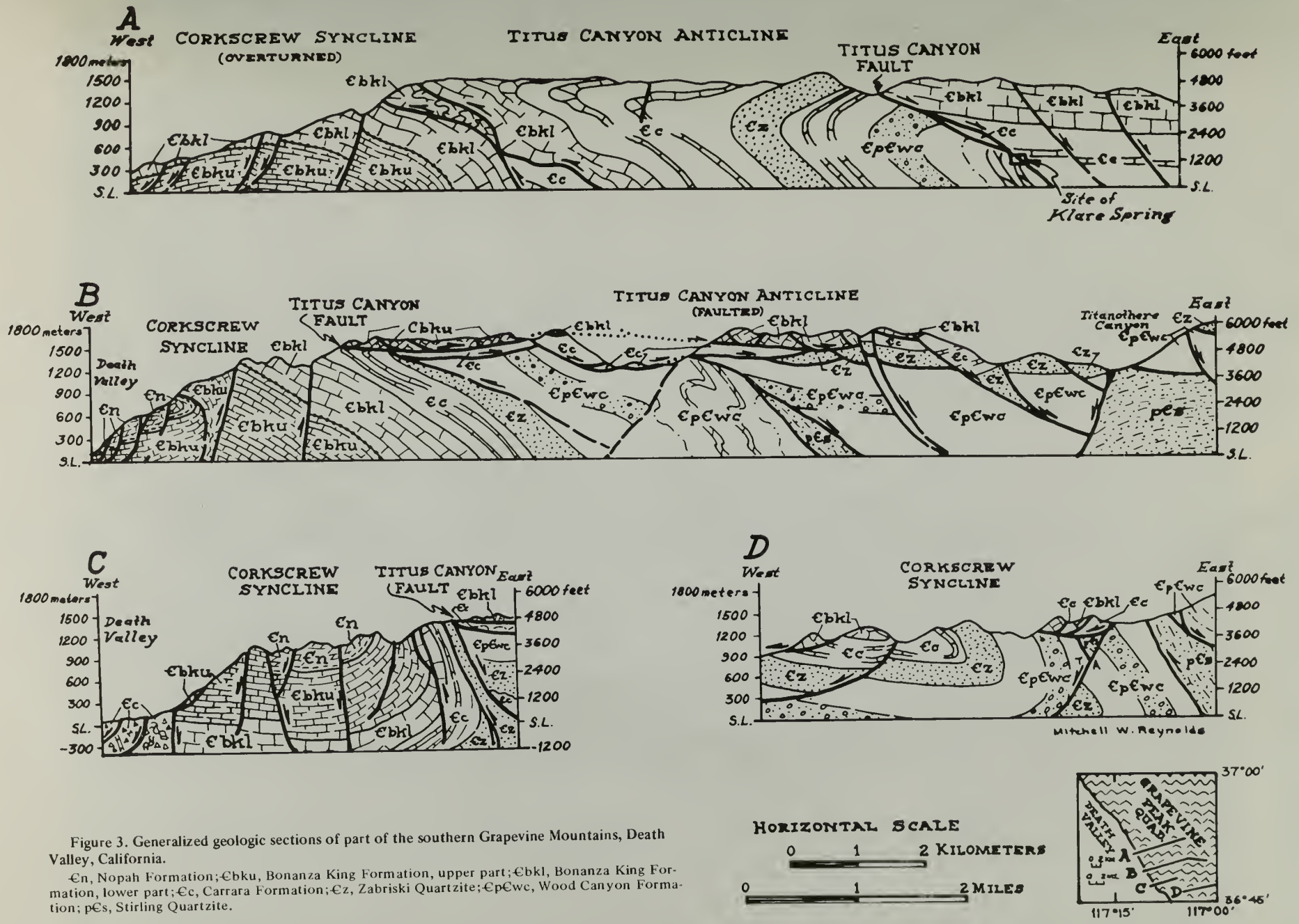


Figure 3. Generalized geologic sections of part of the southern Grapevine Mountains, Death Valley, California.

Cn, Nopah Formation; Cbku, Bonanza King Formation, upper part; Cbkl, Bonanza King Formation, lower part; Cc, Carrara Formation; Cz, Zabriski Quartzite; CpCwc, Wood Canyon Formation; pCs, Stirling Quartzite.

faults flatten eastward to join a single, nearly flat fault called the Titus Canyon fault; it separates the upright but faulted beds from the overturned limb of the recumbent syncline beneath (Fig. 1; B in Fig. 3). That low-angle fault replaces the axial surface of the anticline south of Titus Canyon, whereas north from Klare Spring in the canyon, the fault steepens to become nearly vertical and parallel to the trace of the upper part of Fall Canyon (Fig. 1; A in Fig. 3). Displacement along the vertical segment of the fault is down on the east, and the fault offsets strata as young as early Pliocene.

I have (1969, 1970, 1971) interpreted the flat fault as a lag fault along which the upright limb of the recumbent anticline broke from and lagged behind the developing recumbent fold system. The fault steepens where the recumbent anticline remains intact (A in Fig. 3). Tertiary rocks are absent from the western part of the faulted limb, although east near Leadfield the Titus Canyon Formation is locally preserved in blocks rotated downward toward the principal low-angle fault. These relations, together with the displacement of lower Pliocene rocks along the steep segment of the fault and the style of faulting, all suggest that the recumbent folding and faulting occurred at shallow depths in the crust probably during middle Pliocene time. Evidence further corroborating the young age of the recumbent folding is derived from dating the décollement fault at the base of the fold system as probable middle Pliocene, because that fault offsets strata of early Pliocene age east of Chloride Cliff. The folding was accomplished before late Pliocene and Quaternary time, because beds of fanglomerate of those ages southwest and west of Death Valley Buttes and north of Fall Canyon contain clasts derived from rocks not exposed in the mountains prior to the folding and erosion.

North-trending normal faults on the east flanks of the Grapevine Mountains and northern Funeral Mountains displace the sequence of Tertiary rocks progressively down toward the east. These faults flattened eastward, or locally westward, with depth to follow horizons nearly parallel to bedding in the Cambrian Zabriskie Quartzite or in the upper part of the Wood Canyon Formation (Fig. 2; B in Fig. 3). Displacement on separate faults is as much as 1,000 m (3,200 ft). The offsets suggest that the Cenozoic rocks were extended in a west-northwest–east-southeast direction as they tore from and moved against the subjacent Paleozoic and Precambrian rocks that were being arched in the mountain core.

The abrupt west-southwest and south fronts of the Grapevine Mountains mark zones of normal faulting and warping along which the mountains have risen with respect to Death Valley. Gravity data of Mabey (1963) and the distribution of Tertiary rocks in the southern part of the Grapevine Mountains suggest that vertical separation on the pre-Tertiary surface between the valley and the mountains may be at least 4.3 km (2.7 mi). Fault scarps dip 44° to 75° toward the valley, and antithetic faults dip into the mountain front to break it into a mosaic of strongly fractured small blocks. Warping toward Death Valley accounts for an unknown but substantial part of the structural relief on the range front, for the axial surface of the recumbent syncline of probable middle Pliocene age has been warped 15° down toward Death Valley (A and B in Fig. 3). As a result of both the warping and recumbent folding, beds of the Cambrian Bonanza King Formation at the mouth of Titus Canyon have been rotated through as much as 235° from original horizontality. North of Fall Canyon the apparent vertical separation between Death Valley and the mountain front diminishes and right-lateral strike-slip faulting becomes important within the range. Movement seems to have recurred through latest Tertiary and Quaternary time.

From Death Valley Buttes to Titus Canyon along the steep mountain front, strata in blocks as much as 1.9 km (1.2 mi) across are faulted downward toward the valley along faults that flatten with depth to become nearly horizontal (Fig. 1; C and D in Fig. 3). The flat toes of such faults are commonly exposed at elevations higher than the valley floor. Clearly, these are block-glide landslide structures that formed during or after development of the high relief on the mountain front.

Evidence of Quaternary movement along the northern Death Valley–Furnace Creek fault zone is present almost continuously for 3.2 to 6.5 km (2 to 4 mi) west of the Grapevine Mountains in Death Valley. The linear west front of the Kit Fox Hills, west of Death Valley Buttes (Fig. 1), passes northnorthwest into a linear fault trace along which alluvium and playa sediments are furrowed, and springs emerge. The northern Death Valley highway just south of its junction with the Titus Canyon road rises onto a scarplet in alluvium that varies from 0.6 to 2 m (2 to 7 ft) high; the west side is up relative to the east side. Northwest of Red Wall Canyon an old alluvial fan is displaced right-laterally about 46 m (150 ft) with no significant vertical offset, yet 12 km (7.5 mi) farther northwest, displacement on the fault is about 4.5 m (15 ft) west side up. Near Grapevine Ranger Station, at the north end of the Grapevine Mountains, the east side of the fault is elevated with respect to the west. The linear trace of the fault, patterns of offset of Quaternary deposits, and physiographic features characteristic of strike-slip faulting demonstrate that the fault is nearly vertical and that movement has been in a right-lateral direction. Right-lateral movement on the fault zone, probably occurring largely since middle Cenozoic time, has been variously estimated as being a few miles (Wright and Troxel, 1967, 1970) to scores of miles (Stewart, 1967; Stewart and others, 1968; Burchfiel and others, 1970).

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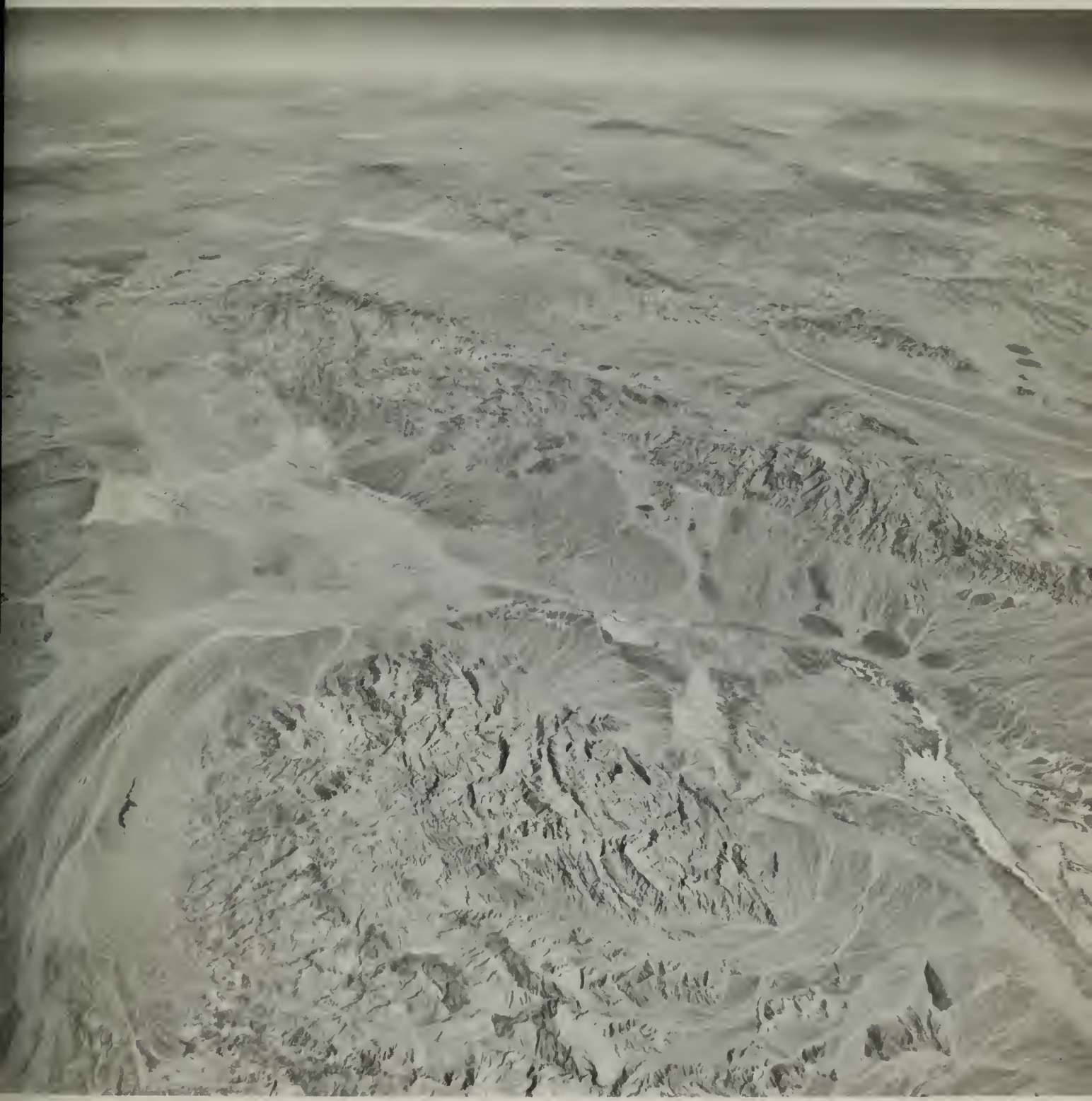

High-altitude oblique aerial photograph to north-northeast from a point approximately above Skidoo, Ponamint Range. Death Valley extends obliquely across center of photograph, Funeral Mountains (right) and Grapevine Mountains (left) lie beyond. The Amargosa River flows southward through the Amargosa Desert from its headwaters area in upper left-center of photograph. The Northern Death Valley–Furnace Creek fault zone forms a modest trace near the base of the fans extending into Death Valley from the Funeral and Grapevine Mountains. Tucki Mountain occupies center foreground. *Photo U.S. Air Force 374L 190, 6 September 1968; courtesy of the U.S. Geological Survey.*

Low-altitude oblique aeriol photograph of Copper Canyon turtlebock in Block Mountoins (see following poper). Floor of east side of Deoth Volley in foreground. *Photo by Lauren A. Wright.*

Geologic Features of the Central Black Mountains, Death Valley, California

James K. Otton[1]

INTRODUCTION

The Black Mountains, which form the eastern margin of the central part of Death Valley, California, contain complex lithologic and structural features that have been difficult to integrate with the geologic framework of the surrounding region. They are underlain mostly by a crystalline complex composed of folded pre-Cenozoic metasedimentary units, in part marble bearing, and of intrusive igneous rocks of Precambrian, Mesozoic(?), and Cenozoic age. Especially abundant in the complex is diorite that postdates the sediments, now metamorphosed, and predates other intrusive bodies that are more acidic and largely to wholly Cenozoic in age. Overlying the complex, generally with fault contact, are structural units composed variously of slightly metamorphosed Precambrian and Paleozoic sedimentary rocks and Cenozoic sedimentary and volcanic rocks. These constitute a relatively thin, discontinuous, and faulted cover of the complex.

The following is a report on the progress of a mapping project in the southwestern Black Mountains. The area (Fig. 1) includes the southwestern quarter of the Funeral Peak and the southeastern part of the Bennetts Well 15′ quadrangles. In this area, the igneous and metamorphic complex—a dislocated cover of Cenozoic rocks—and two of the peculiar domical features known as turtlebacks are well exposed. Although many questions concerning the geologic history of the Black Mountains remain unanswered, the mapping to date has documented several features relevant to this history and the evolution of the region.

The general geologic features of the Black Mountains have been shown on several published maps, namely those of Curry (1954), Noble and Wright (1954—generalization of Curry's map), Drewes (1963), the Death Valley sheet of the State Geologic Map of California (scale 1:250,000), and Wright and Troxel (1973). Curry (1954) outlined the rock units and structural features along the western margin of the north and central Black Mountains and was the first to describe the turtleback surfaces. He concluded that they were parts of a folded and eroded thrust fault, and he correlated the fault with the Amargosa thrust fault that Noble (1941) had identified in the southern Black Mountains.

Drewes, working in the Funeral Peak quadrangle and in an adjacent part of the Bennetts Well quadrangle (Drewes, 1963), also described the rock units and structural features; he concluded that the turtleback surfaces and the Amargosa thrust were separate features. He favored the hypothesis that the turtlebacks were produced by differential erosion during Tertiary time of a terrane of folded metamorphic rocks, producing an undulating surface on which late Tertiary rocks were deposited and from which they slid along normal faults.

The geologic features of the area in Figure 1 present numerous avenues of investigation, including the following: (1) the stratigraphy and age of the sedimentary rocks now metamorphosed; (2) their metamorphic history and the role of the various intrusive events in the metamorphism; (3) the nature and time of folding of the metamorphic rocks; (4) the age and mode of emplacement of the diorite and the bodies of younger igneous rock; and (5) the evolution of the fault pattern, including the faulted surfaces between the complex and the bodies of rock that overlie it.

LITHOLOGIC AND STRUCTURAL FEATURES OF THE METASEDIMENTARY ROCKS

The metasedimentary rocks of the crystalline complex can be subdivided into two gross units. The stratigraphically lower unit is composed of augen gneiss, biotite schist, amphibolite, biotite-hornblende gneiss, quartz-feldspar gneiss, and gneissic granite. This unit lithologically resembles the older Precambrian complex extensively exposed in the more southerly parts of the Black Mountains and is here correlated with it. The stratigraphically higher unit is carbonate bearing and consists variously of massive light-brown to gray dolomitic marble; well-bedded tan stromatolite-bearing dolomitic marble (Fig. 3) with cherty lenses and nodules; tan dolomitic marble containing thin layers of hornfels; chert; quartzite; and light-gray limestone. It displays a maximum thickness of about 600 ft.

Wherever the two units are in contact, the carbonate-bearing unit overlies the generally gneissic unit. In addition to their lithologic differences and stratigraphic relations, differences in style of deformation and metamorphism are distinctive. The gneissic unit is characterized by variously oriented mesoscopic, isoclinal, locally recumbent folds and by textures attributable to shearing. The carbonate unit, on the other hand, is characterized by upright slightly overturned macroscopic disharmonic folds that consistently plunge northwestward. The older unit is pervasively metamorphosed to lower amphibolite or upper greenschist facies. The metamorphism of the carbonate-bearing unit ranges widely in intensity and apparently is related to distance from contacts with the diorite. At localities farthest from diorite contacts, chert bodies are devoid of prograde metamorphic minerals and of evidence of recrystallization. Chert nodules in exposures of the carbonate-bearing unit close to contacts with diorite are altered to talc, epidote, and (or) garnet and have been recrystallized. Finally, the contact between the carbonate-bearing unit and the older complex is discordant with foliation in the complex.

Between Copper Canyon and Sheep Canyon (Fig. 1), the older Precambrian schist and gneiss form an elongate dome that is mantled by a thin, discontinuous sheath of the carbonate-bearing unit. The morphology of this dome is displayed in a structure contour map of the contact between the gneissic unit and the carbonate-bearing unit (Fig. 2). The dome is steepest on its southwestern flank, which is marked by a broad fault zone containing abundant large fragments of carbonate rock. Diorite lies on the southwestern downdropped side of this fault. The type turtleback (Curry, 1954) is located on the northwestern nose of this dome. Geometrically, this structure resembles a mantled gneiss dome (Eskola, 1949), in that it consists of a core of gneiss that underlies a cover of younger, structurally discordant metamorphosed rocks.

The older Precambrian rocks mapped in the Mormon Point area are also mantled by a sheath of carbonate-bearing rocks. The sequence preserved there is thicker than on the other dome and obscures most of the older Precambrian units. The exposures of carbonate-bearing rocks terminate, however, along

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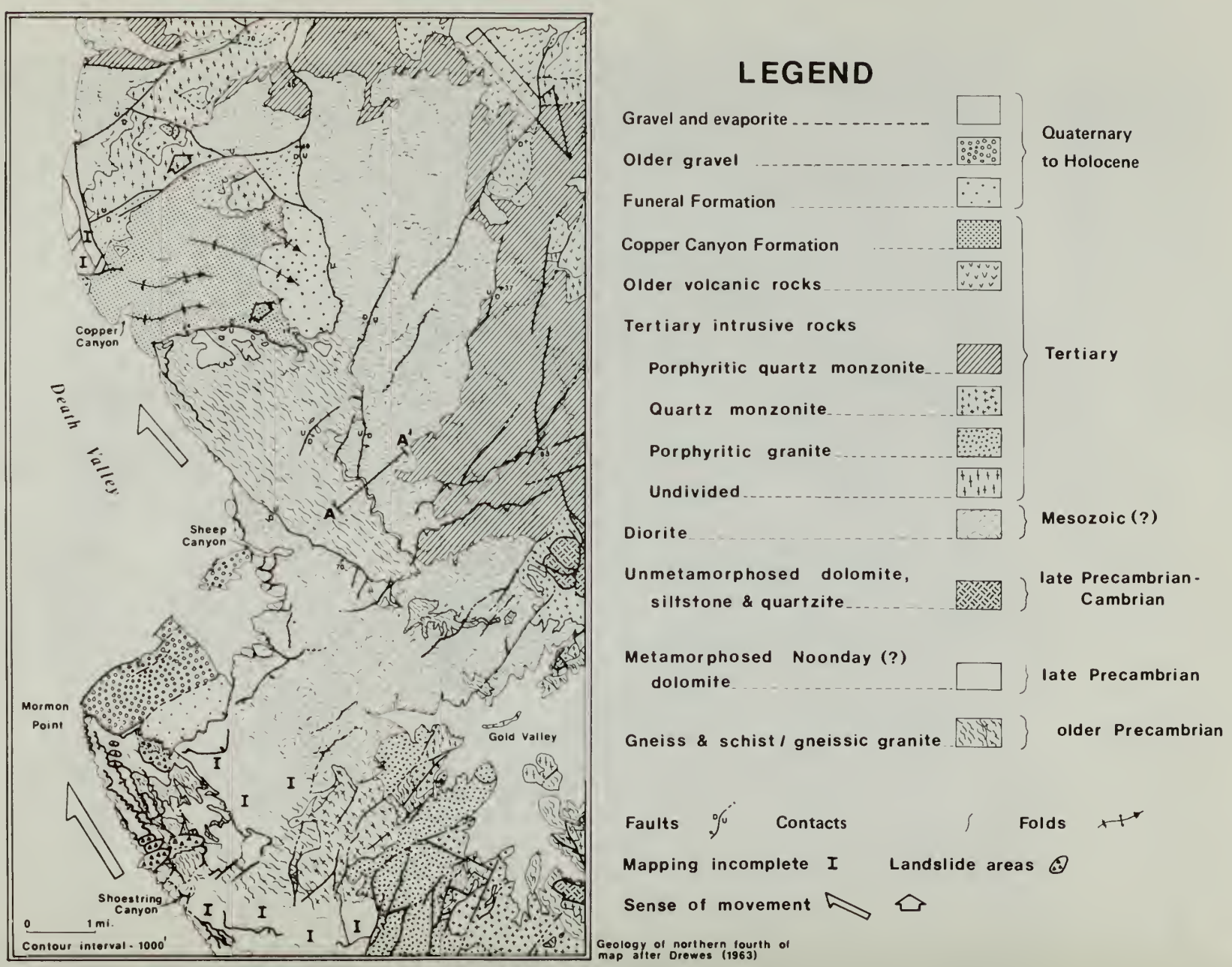

Figure 1. Generalized geologic map, central Black Mountains, Death Valley, California.

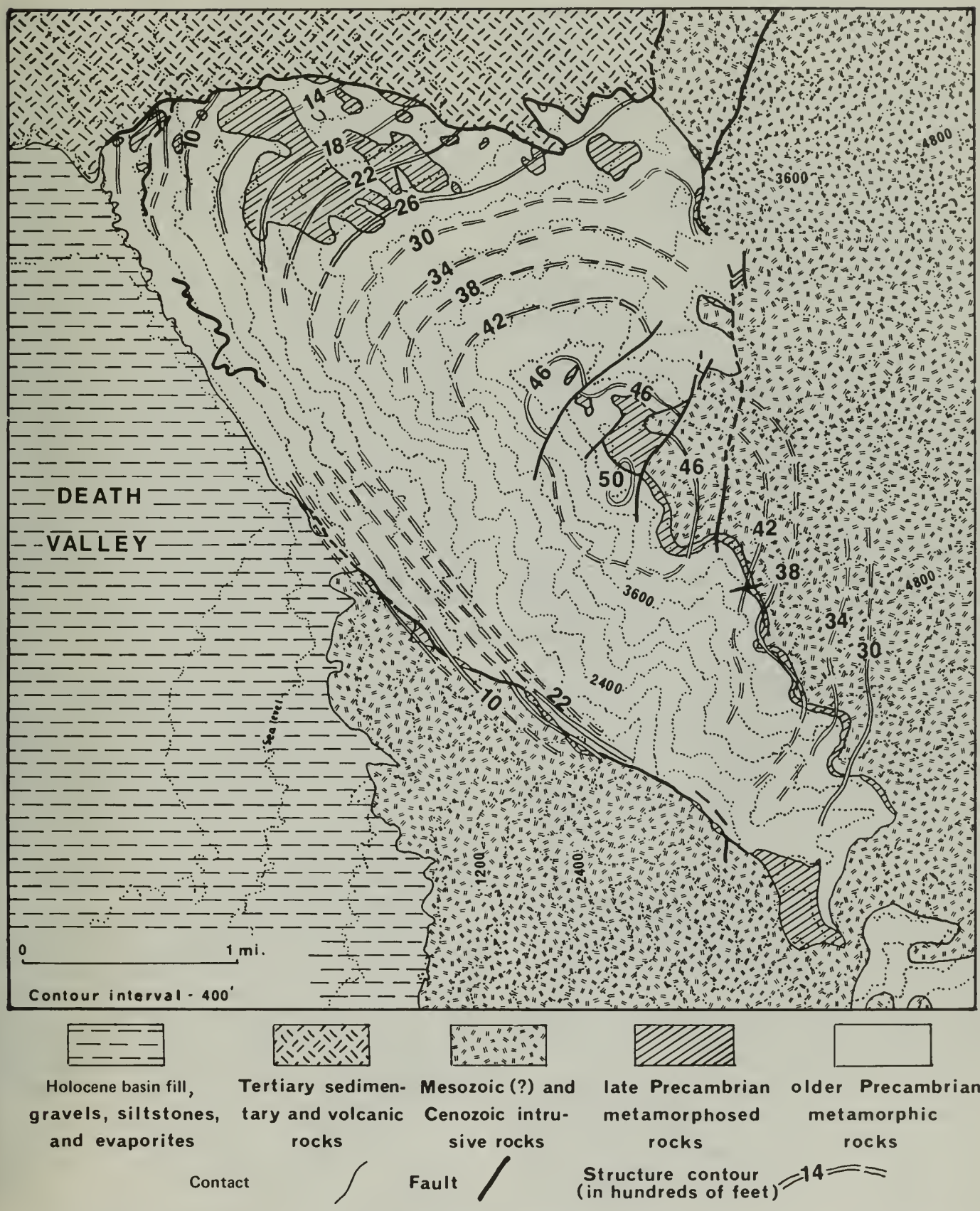

Figure 2. Generalized geologic map of a part of the Black Mountains showing structural contours on contact between older Precambrian and late Precambrian units.

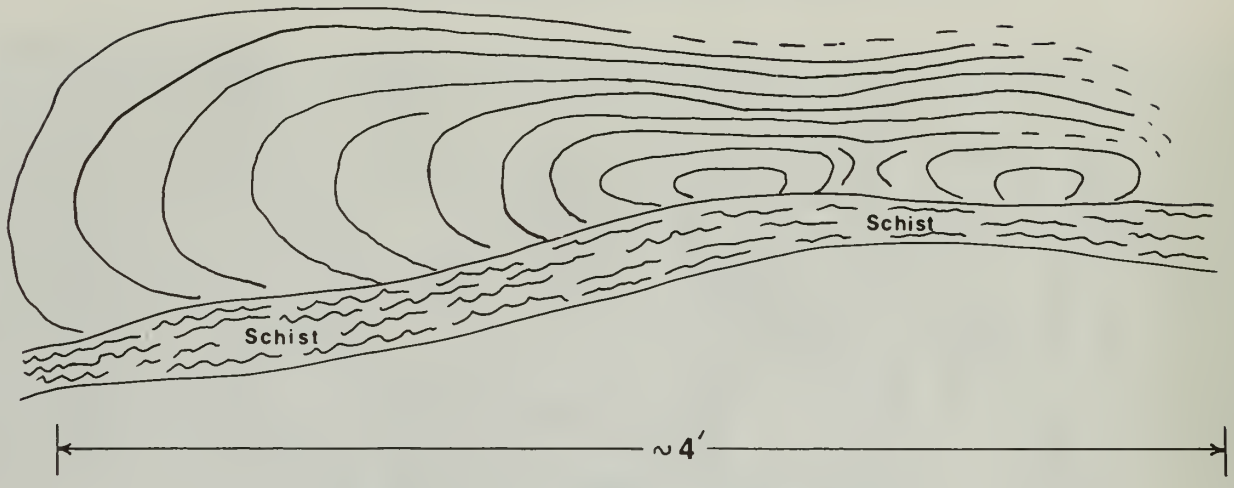

Figure 3. Sketch of stromatolites in carbonate, Mormon Point.

a reverse fault on the south side of Shoestring Canyon (Fig. 1). Although there are extensive exposures of older Precambrian schist and gneiss south of this fault, no mantling carbonate-bearing sequence is exposed. The unconformity beneath the carbonate-bearing rocks is folded in the Mormon Point area, but the shape of that surface cannot be completely reconstructed. Minimum relief on the surface is 1,200 ft.

CORRELATION OF THE MARBLE WITH OTHER CARBONATE UNITS OF THE REGION

If the diorite is Precambrian in age and is older than later Precambrian Pahrump Group and younger rocks that are extensively exposed in the southern part of the Black Mountains (Drewes, 1963), then the marble must be older than the Pahrump Group and composes a unit unexposed in the pre-Pahrump terrane of the Death Valley region.

If the diorite is Mesozoic in age, the marble may be correlated: (1) with carbonate units of the Pahrump, (2) with the Noonday Dolomite and (or) the Johnnie Formation, (3) with other carbonate units of later Precambrian or Paleozoic age, or (4) with carbonate units of Triassic age.

Correlation with the Noonday Dolomite or the Johnnie Formation is compatible with the paleogeographic reconstructions of Wright and Troxel (1966, 1967), with the lithologic descriptions of the Noonday by Williams and others (1974) and Williams (1974, personal communication) and with the lithologic descriptions of the Johnnie Formation by Benmore (1974). In wright and Troxel's reconstructions, the Mormon Point and Copper Canyon–Sheep Canyon areas are part of the Nopah Upland, an area from which Pahrump Group sedimentary rocks were eroded in late Pahrump and pre-Noonday time. It is thought that parts of the Noonday Dolomite and the Johnnie Formation then transgressed and possibly blanketed the post-Pahrump terrane (Fig. 5). If so, then these carbonate units could be rocks of the Noonday Dolomite or the Johnnie Formation metamorphosed near the diorite body, if it is younger than the Noonday.

In general aspects, much of the carbonate sequence of Mormon Point resembles the shelf facies of the Noonday Dolomite described by Williams and others (1974). The massive carbonate rocks and the thinly-bedded carbonate rock with hornfels lenses may be the metamorphosed equivalent of massive domal carbonate units and thinly-bedded carbonate units with cryptalgal structures in the Noonday shelf facies. Moreover, sharp lateral thickness variations noted by Williams and others (1974) also occur in carbonate units at Mormon Point.

Parts of the sequence contain rocks similar to parts of the Johnnie Formation. This includes a ferroan dolomite and cyclic units consisting of clastic carbonate and fine-grained siliceous material. The clastic carbonate is cross-bedded (cross-beds 6 in. to 18 in. high) and exhibits a bimodal rosette that suggests a west-dipping paleoslope and a tidal environment.

Correlation with carbonate units in the Pahrump Group (possibly preserved as in Fig. 4), namely a carbonate member in the Crystal Spring Formation (Roberts, this volume) or with the overlying Beck Spring Dolomite also is a possibility, however, the marble of the Black Mountains is unlike the carbonate units of the Pahrump or any of the younger units in the late Precambrian, Paleozoic, and Mesozoic terranes of the Death Valley region.

DIORITE BODIES

The pluton or plutons of diorite that intrude the metamorphic rocks of the Copper Canyon–Sheep Canyon and Mormon Point–Smith Mountain areas extend discontinuously from the southwestern flank of Smith Mountain to Badwater, 6 mi north of the present map area, a distance of 15 mi (Drewes, 1963, Pl. 1). It comprises more than half the mapped crystalline rocks of the central Black Mountains. The original shape of the diorite pluton(s) is difficult to document because many of its contacts are faulted or are intrusive contacts with later granitic bodies. Near the Copper Canyon turtleback, however, the diorite occurs as a sill that follows closely the contact between the two units of metamorphic rocks (Fig. 6). In the area of Mormon Point, the intrusive relations are less simple.

Diorite contacts in the Copper Canyon–Sheep Canyon area are sharp and can be located within a few feet. Near Mormon Point the contacts are gradational over a few hundred to several hundred feet. Foliation in the older Precambrian gneiss becomes indistinct, and the gneiss commonly develops feldspar porphyroblasts near the diorite contact.

Xenoliths of metamorphic rocks in the diorite are common, particularly in a zone that extends southeast from the central part of Sheep Canyon into Gold Valley. The xenoliths consist mostly of the older Precambrian gneiss, but carbonate quartzite and schist form xenoliths on the north side of Gold Valley.

The petrology of the diorite, as described by Drewes (1963, p. 8–9), consists mostly of hornblende and plagioclase with minor or accessory proportions of biotite, clinopyroxene, sphene, titaniferous magnetite, and apatite. The texture is hypidiomorphic granular in most occurrences. In others, hornblende exhibits ophitic or poikilitic textures. Plagioclase is commonly poikilitic. Drewes stated that some of the hornblende may represent secondary growth from pyroxene and

that some of the plagioclase may also be secondary. He added (1963, p. 9–10):

> . . . This replacement is the chief metamorphic process that has affected the original diorite or gabbro; the metamorphism could have occurred at the same temperature and pressure as the metamorphism of the (older Precambrian) metasedimentary rocks.
>
> The metadiorite is younger than the metasedimentary rocks, and it is possibly also younger than the folding of the metasedimentary rocks, for it is virtually unfoliated and it intrudes and appears to truncate the foliated rocks. In degree of metamorphism it has a far greater resemblance to the older Precambrian rocks of the Funeral Peak quadrangle than to the Pahrump series of late Precambrian age; therefore the metadiorite is tentatively assigned to early Precambrian.

The textures that Drewes attributed to metamorphism could also be ascribed to deuteric alteration of the intrusive body during cooling. Reliable radiometric dates have yet to be obtained for the diorite. However, it resembles diorite that is clearly Mesozoic in age in other bodies in the Death Valley region. These occur in the Avawatz Mountains (B. W. Troxel and L. A. Wright, personal commun.) and the southern part of the Panamint Range (Johnson, 1957). A post-Precambrian age for the diorite is also suggested by the observation that it extends south into an area where the Precambrian gneiss is cut by numerous diabase dikes that were emplaced in Pahrump time. The diorite there is not intruded by diabase, although none of the diorite contacts has been observed to actually show crosscutting relationships with the diabase. Moreover, unlike the metasedimentary units, the diorite is devoid of pervasive foliation. That it postdates the folding of the metasedimentary rocks is indicated by mafic dikes that extend into the folded rocks from the pluton(s) of diorite, without evidence of folding. The diorite crosscuts faults which, in turn, cut exposures of the folded carbonate sequence. Thus, the diorite is believed to represent a much later event related to the Mesozoic plutonism so common in the region west and south of Death Valley.

SILICIC PLUTONS AND DIKES

The igneous bodies that intrude the diorite and the metasedimentary units range in composition from granite to monzonite. The mappable igneous bodies comprise a pink porphyritic granite, two types of quartz monzonite, three types of porphyritic quartz monzonite or latite, red porphyritic rhyolite, and andesite dikes.

To completely reconstruct the sequence of intrusive events through cross-cutting relationships is difficult. The pink porphyritic granite and the porphyritic quartz monzonite intrude the diorite but not each other. A gray phase of the quartz monzonite clearly intrudes the pink porphyritic granite, as demonstrated by xenoliths of the latter in the monzonite. A white phase of the quartz monzonite intrudes the diorite, but its contact with the porphyritic quartz monzonite north of Sheep Canyon is unexposed.

The red porphyritic rhyolite that forms plugs and the andesite that occurs as dikes probably represent the latest events. The andesite dikes are abundant on the south side of Gold Valley and Smith Mountain where they intrude the gray phase of the quartz monzonite. One andesitic plug intrudes the porphyritic quartz monzonite near the northeastern corner of the area. The red porphyritic rhyolite intrudes both the porphyritic quartz monzonite and the quartz monzonite at various localities.

The intrusive contacts between these various igneous bodies and their host rocks are commonly gradational over distances of hundreds of feet or more (Fig. 7).

The larger plutons commonly are bordered by an outer zone where only a few dikes penetrate the host rock, an intermediate zone where the number of dikes increases to form an anastomosing network in which the host rock "floats" as xenoliths, and finally an inner zone where the number, size, and volume of xenoliths approaches zero. The anastomosing dikes are largely horizontal, and xenoliths commonly maintain a tabular, subhorizontal dimensional orientation. On the other hand, many contacts are sharp. The pink porphyritic granite exhibits a sharp subhorizontal contact with the older Precambrian rocks on the southwestern flank of Smith Mountain.

Dike swarms and dike systems that are compositionally related to the silicic plutons extend beyond the margins of those intrusive bodies. Many of the dike systems show no systematic orientation, while others show a strong preferred orientation. One system of dikes occupies the crest of a ridge that lies between Sheep Canyon and Copper Canyon. This dike system consists of gray porphyritic quartz monzonite similar to one of the porphyritic intrusive phases a few miles to the east. The dikes are remarkably planar; they are all subvertical, and all but two strike approximately N. 30° E. They are spaced 200 to 800 yd apart, and their thicknesses reach 75 ft and

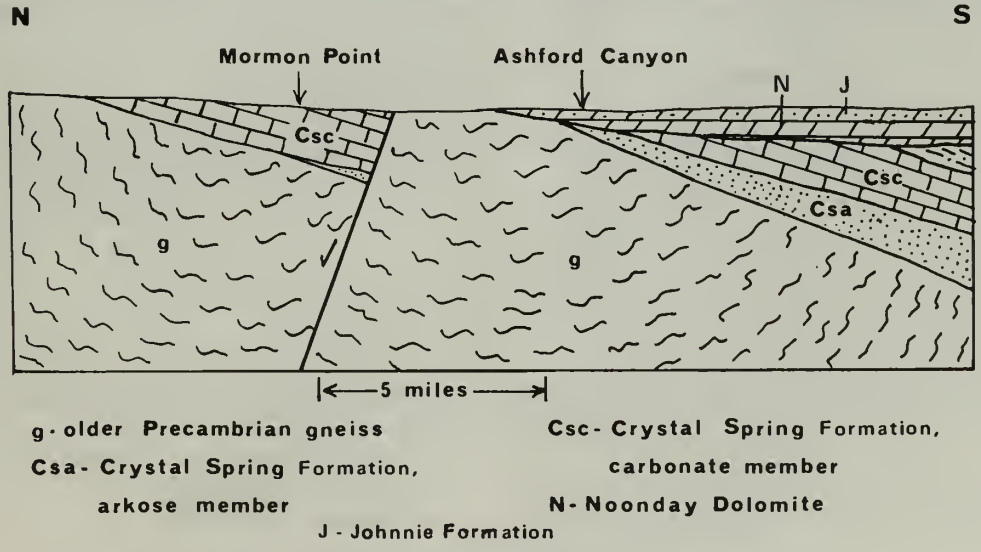

Figure 4. Cross section of hypothetical late Johnnie Formation Black Mountain terrane showing preservation of carbonate member of Crystal Spring Formation adjacent to a fault.

lengths over 2 mi. This planarly oriented system of dikes is restricted to the older Precambrian rocks of the ridge. The dikes terminate to the southwest against the range front fault and to the northeast against the turtleback fault. Some of the dikes cross the intrusive contact into the diorite, but where they do, they change orientation.

A second system of dikes occupies the northerly and northwesterly faces of a ridge of diorite just to the south of the mouth of Sheep Canyon. Dikes in this system are essentially planar but interfinger more than the dikes of the previously described system. They strike N. 75° E. and dip 35° to 45° to the southeast. These dikes also consist of porphyritic quartz monzonite. They are more closely spaced than the other system—as close as 100 ft or less.

Most of the silicic dikes and all of the silicic plutons of the region lack a visible foliation. One swarm of dikes, composed of gray aphanitic to slightly porphyritic rock, shows well-developed layering. Layers are distinguished by varying shades of gray, or gray and reddish gray. Layer thickness varies from a few millimeters to a few centimeters. The layering is interpreted as flow layering because isoclinal folds symmetric with dike margins are common in the layers. In larger dikes, the central portions commonly are unlayered.

The age of these igneous bodies is uncertain. Drewes (1963) reported two lead-alpha age determinations on zircon in the monzonitic rocks, 45 ± 10 and 30 ± 10 m.y. These determinations, made in 1959, are probably too old by a large factor (Rose and Stern, 1960). Fleck (1970) reported K-Ar whole-rock age dates on the older volcanic rocks of the region of 6 to 8 m.y. Because these units were deposited on or were involved in faulting that cuts the monzonitic rocks, these dates probably can be considered a minimum age for the monzonites. The time of intrusion for these rocks seems best placed between 10 and 15 m.y. ago.

TERTIARY AND QUATERNARY SEDIMENTARY AND VOLCANIC ROCKS

Drewes (1963) described in detail the lithology of the Tertiary and Quaternary rocks of the region, the older volcanic rocks, the Copper Canyon Formation, the Funeral Formation, and gravel of various ages. The older volcanic rocks include rhyolite and rhyodacite flows and tuff with interbedded tuff-breccia and agglomerate and minor volcanigenic sediments. The Copper Canyon Formation comprises over 10,000 ft of reddish conglomerate, yellowish-gray siltstone, and evaporites and intercalated basalt. Based on mammalian fossils, it is thought to be Pliocene in age (Curry, 1941). The Funeral Formation consists of tan conglomerate and diorite megabreccia.

STRUCTURAL FEATURES

The structural features of the central Black Mountains appear to have developed in two or more stages. The latest stage has produced faults which are compatible with a northwesterly crustal extension of the Black Mountains–central Death Valley terrane. The earliest stages involved the development of the major structural features of the crystalline terrane before extension began. This earlier deformation and intrusion resulted in a tectonic fabric that controlled the faulting during the northwest crustal extension of the later stage. The silicic plutons may mark the beginning of the extensional stage.

The features of the latest stage include many northeast-trending intermediate to high-angle normal faults with displacements generally within the range of a few tens of feet to a few hundred feet. The features of the latest stage mainly consist of faults that are either clearly normal or seem best interpreted as normal in sense of movement. The most abundant of the well-defined normal faults strike northeast and ordinarily show displacements measurable in tens or hundreds of feet; some of the displacements are 1,000 ft or more. One of the larger of the northeast-trending faults bounds the western side of the depression that contains the Copper Canyon Formation (Fig. 1). The faults that derive their configuration from earlier structural features and that are also interpreted here as being normal include the turtleback faults at Mormon Point and Copper Canyon and extensions of the turtleback faults. Linear features, including slickensides, fault mullions, and drag folds, on the surfaces of these faults indicate that the downward movement of the hanging wall has been in a northwesterly direction.

Apparently related to the normal faults are two high-angle faults that strike northwest and can be traced for 5 mi or more (Fig. 1). The linear features of movement on the planes of these faults indicate movements in a normal sense and also in a right-lateral oblique-slip sense. One of these faults terminates along strike in the lower part of Sheep Canyon.

The nature and distribution of the faults of the Black Mountains escarpment support the long-held view that the Black Mountains front is essentially an irregular fault scarp developed during the latest stage of deformation. It consists of

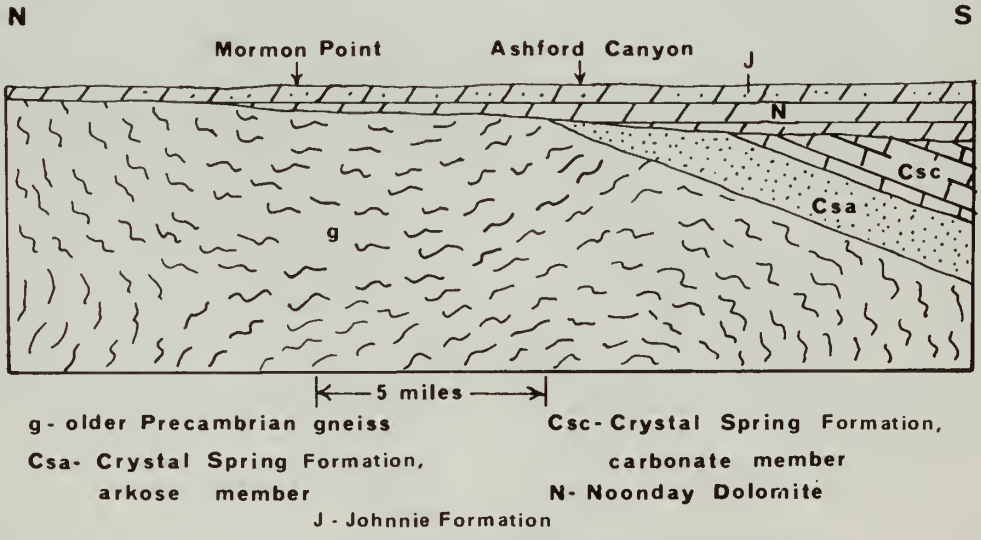

Figure 5. Cross section of hypothetical late Johnnie Formation Black Mountain terrane showing onlap of Noonday Dolomite, in turn overlain by Johnnie Formation.

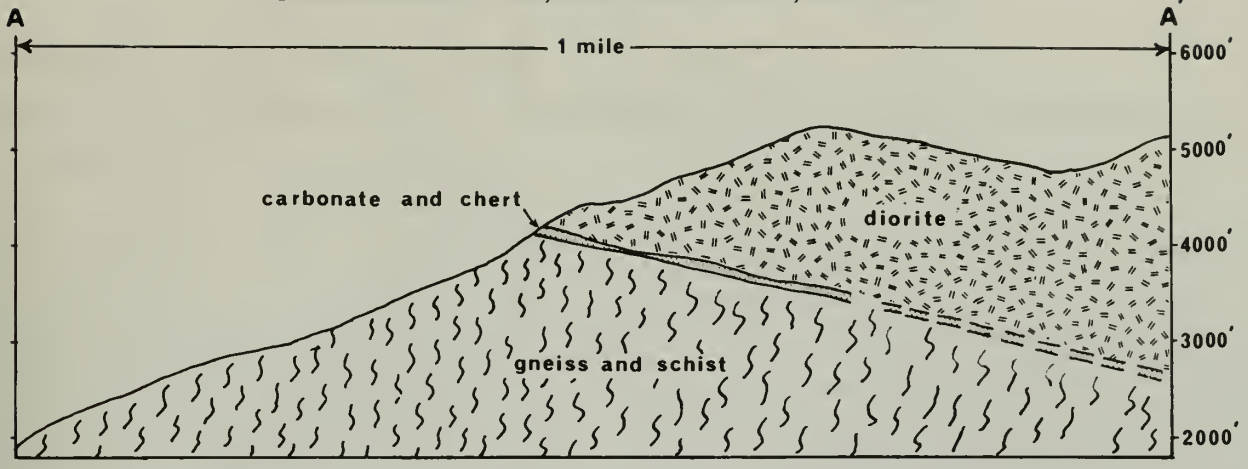

Figure 6. Cross section of ridge between Sheep Canyon and Copper Canyon showing diorite contact.

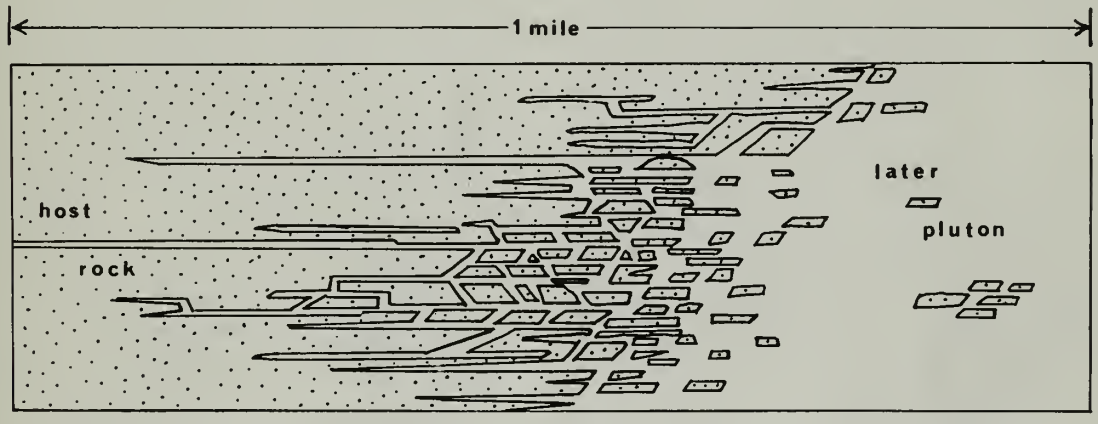

Figure 7. Schematic cross section of gradational contact between silicic plutons and host rock.

northwest-trending en echelon faults with linking northeasterly irregular zones. The irregular zones, however, contrary to previously expressed interpretations, are also related to crustal extension. They coincide geometrically with weak horizons (the carbonate-bearing units) in the early fabric of the crystalline rocks. This suggests that the orientation and position of these zones was controlled by zones of weakness in the inherited fabric.

This view of the Black Mountains front has recently been expressed by Wright and others (1974, reprinted in this volume) and is compatible with the pull-apart model (Fig. 8A) for the origin of the central segment of Death Valley (Burchfiel and Stewart, 1966). Figure 8B is a modification of that model and shows the structural features of the east wall of the central part of Death Valley in greater detail.

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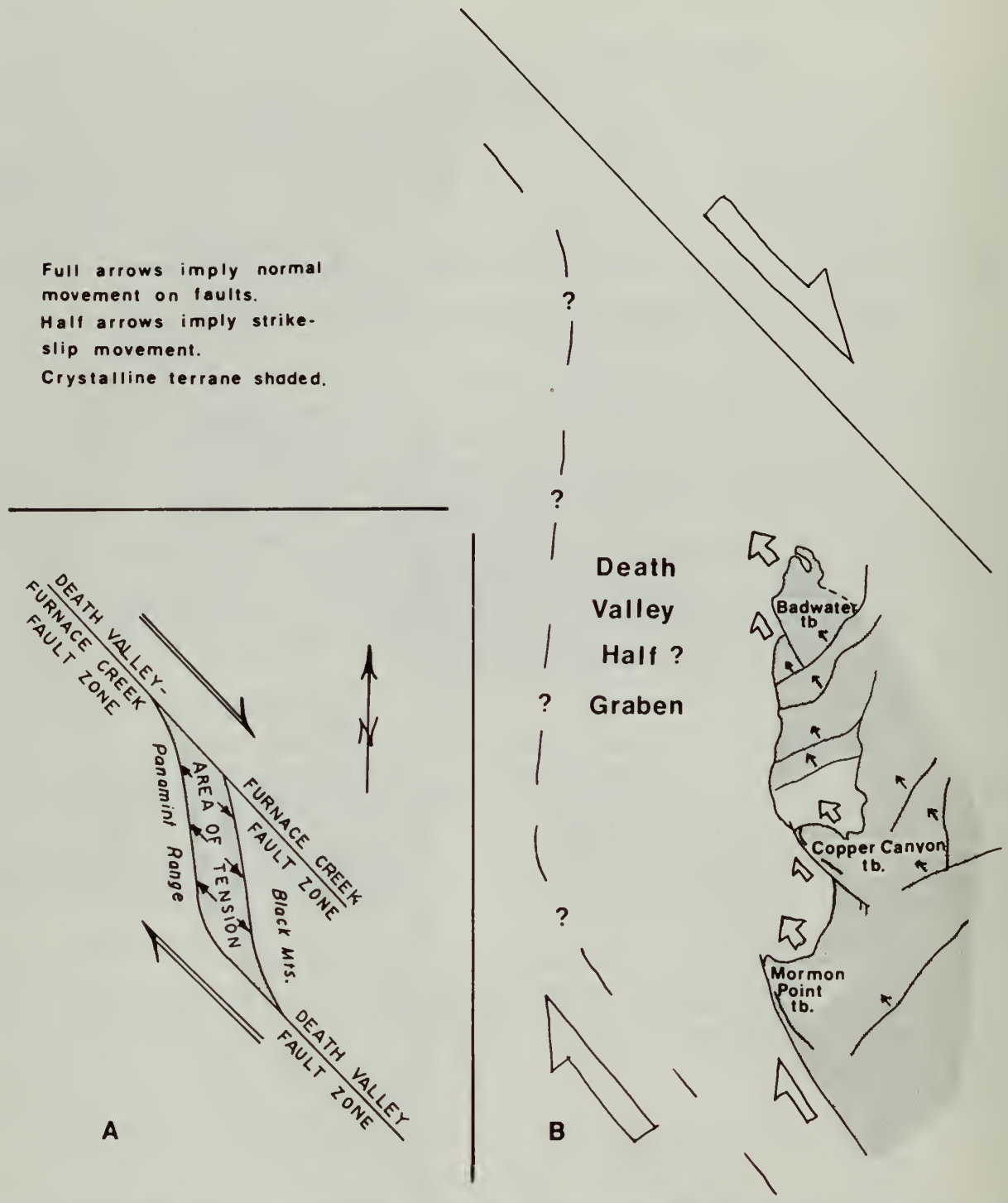

Figure 8. A. Diagrammatic map showing interpretation of strike-slip movement and area of tension (from Burchfiel and Stewart, 1966). B. Detailed map of east wall of area of tension (generalized from Death Valley sheet, Geologic Map of California).

Stratigraphy and Depositional Environments of the Crystal Spring Formation, Southern Death Valley Region, California

Michael T. Roberts[1]

INTRODUCTION

This report is a summary of a study of the lower and middle part of the Crystal Spring Formation in the southern Death Valley region begun in 1972. The lower three members of the formation have been studied throughout the region; the dolomite and algal members were investigated at selected localities. The upper two members and the diabase sills in the formation were not studied, and data pertaining to them in this report are primarily from Wright (1968).

The late Precambrian Crystal Spring Formation is exposed within a 75 by 25 mi (121 by 40 km) belt that trends approximately east and extends from the Kingston Range to the southern Panamint Range. The Crystal Spring is the lowest of three formations of the Pahrump Group. It rests with profound unconformity on older Precambrian metamorphic and igneous rocks, and it is successively overlain by the Beck Spring Dolomite and the Kingston Peak Formation. The older Precambrian complex has been dated at 1,700 m.y. (Wasserburg and others, 1959). Diabase sills within the Crystal Spring have been correlated with sills 1,200 m.y. old in Arizona on the basis of their similarity in composition (Wrucke, 1972). The Crystal Spring Formation contains *Baicalia* stromatolites (Howell, 1971); this suggests a middle Riphean age (1,350 to 950 m.y.) for the algal member of the formation. The algal member also contains *Conophyton* stromatolites. Raaben (1969) suggests that the *Baicalia-Conophyton* association is typically middle Riphean. The age may be, then, between 1,350 and 1,200 m.y.

The Crystal Spring Formation is divisible into seven sedimentary members (Fig. 1) and contains diabase sills that range in thickness from a few tens of feet to about 1,500 ft (457 m). The sills are not shown in Figure 1 in order to better show the relations of the sedimentary members. The lower four members, redefined below from those of Wright (1968), are each remarkably persistent throughout the region, varying in thickness but not markedly in lithology. The thickest diabase sill generally occurs within the basal part of the dolomite member, but sills also occur higher in the section in several areas, and feeder dikes are exposed locally. Contact metamorphism has altered the rocks for a distance, generally, of 100 to 150 ft (30 to 46 m) on both sides of the sill. Below the sill, in most exposures, are hornfelsic calc-silicate rocks (the "upper quartzite member" of Wright, 1968, here included with the dolomite member) which were originally impure carbonate rocks rich in quartzose sand and shaly strata. In most areas above the sill, dolomite beds have been altered to talc-tremolite rocks, which are extensively mined in the area (Wright, 1968).

The upper part of the dolomite member and the upper three members (Fig. 1) are persistent in northern and central exposures but apparently show marked facies changes to the south. The southern facies are restricted in outcrop to the Silurian Hills and northern Avawatz Mountains. The exact relation of the northern to southern facies is unknown, because they are separated by a 5- to 15-mi (8- to 24-km) interval in which the Crystal Spring is not exposed, and, also, because of a lack of detailed data on the upper members.

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ARKOSE MEMBER

The arkose member of the formation is here defined as the arkosic sandstone and conglomerate between the basal unconformity and, generally, the top of the uppermost layer of conglomerate. The upper boundary is generally marked by the upward change in color from mainly gray and green to the red or purple of sandstone of the overlying member. In some areas—for example, the Owlshead Mountains—the uppermost conglomerate is absent and the top is defined on the basis of bedforms and color change. In other areas, such as the southern Panamint Range, the conglomerate is present, but the color change occurs within rather than above it. Preliminary petrologic study indicates that the boundary also separates more feldspathic strata from overlying less feldspathic strata, although the change may be gradational.

The member is composed of about 60 percent sandstone and 40 percent conglomerate. Thin beds of shale, siltstone, chert, and carbonate are very subordinate and occur within the finer-grained parts of the member. Thin sections of the member (36 thin sections, 500 points counted per thin section) indicate an average composition of 35 percent quartz grains (mainly undulatory extinction), 22 percent matrix (mainly illitic material), 20 percent rock fragments (15 percent polycrystalline quartz, 5 percent granitic), 17 percent feldspar (albite, microcline, orthoclase), and 6 percent accessory minerals (calcite, opaque minerals, mica, heavy minerals). Much of the matrix appears to be authogenic and derived from the feldspars. "Ghost" grains of feldspar are abundant in the matrix, and the feldspar displays all stages of alteration to illitic material. Consequently, the original feldspar content of the member may have exceeded 30 percent. The rock fragments, heavy minerals, and abundance of alkali feldspar indicate a source containing granitic igneous and metamorphic rocks. The basal and uppermost beds of the member tend to be more quartzose than the rest of the member, trending toward feldspathic sandstone in composition. Variations in quartz/feldspar ratios in the member seem to be related to differences in degree of alteration of feldspars in the sediment and to changes in weathering or composition in the source area, rather than to differences in depositional environment.

Thickness trends of the arkose member and the initial geometry of the depositional basin of the formation are shown in Figure 2. The arkose member contains a basal conglomerate unit as much as 27 ft (8 m) thick, which locally pinches and swells, but which appears to increase in thickness from about 4 ft (1.2 m) to more than 20 ft (6 m) east-southeastward. It consists of pebbles and cobbles composed mainly of metaquartzite and white quartz in a matrix of very coarse feldspathic sandstone. Exposures of this conglomerate in the central part of the region display a clast imbrication. Measurements of clast orientation indicate southerly flowing currents. The basal conglomerate is overlain in many areas by conglomeratic strata (described below) with cross strata that show southerly current directions; it is probably fluvial in origin. In the eastern Kingston Range, the basal conglomerate appears to be part of a fining-upward fluvial sequence. The basal conglomerate contains scour-and-fill structures and potholes at its base. However, in some areas—for example, the southern Ibex Hills—the conglom-

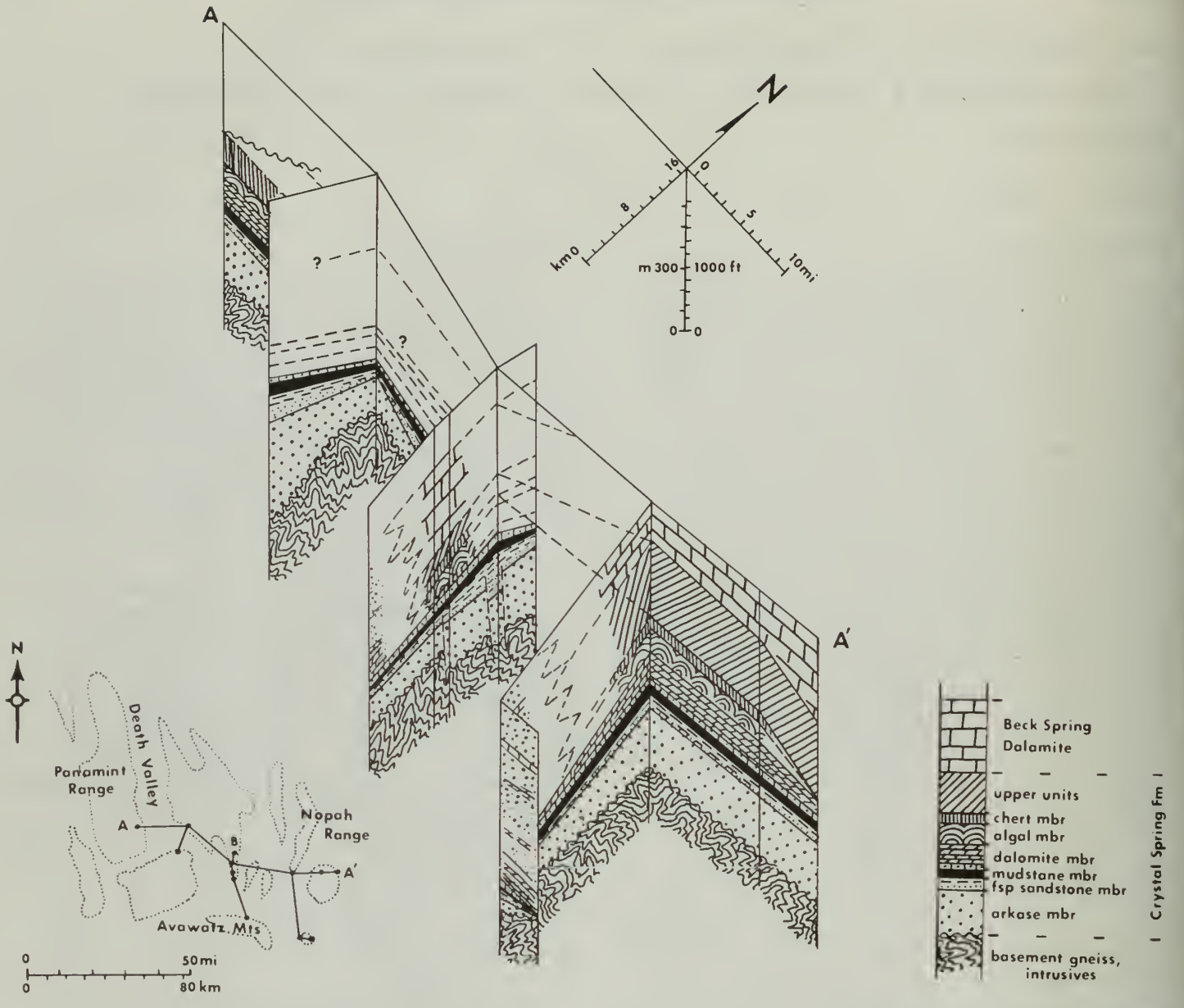

Figure 1. Fence diagram of Crystal Spring Formation. Vertical exaggeration x 26. Location of measured sections (vertical lines of diagram) shown as dots on inset map. Generalized stratigraphic column inset represents northern lithologies. Lithologic changes to south are shown diagrammatically on three southernmost sections. Data for chert, upper units, and Beck Spring Dolomite from Wright (1968). Data for Silurian Hills section in part from Kupfer (1960).

erate is overlain by cross-stratified tidal sandstone beds, which show northerly current directions (described below). The basal conglomerate thus may be the product of complex nearshore processes during the initial transgression over the area. It may, in part, represent a transgressive, cobble beach deposit. The conglomerate appears to have been derived primarily from the erosion of underlying rocks. In the southern Panamint Range, for example, it contains clasts of schist and vein quartz identical with the underlying rocks. The metaquartzite of the clasts is unlike any known older rocks exposed in the region. They may represent redistributed clasts from gravel that formerly overlay the basement or were transported directly from a distant source.

The rest of the member is a cyclical sequence of conglomerate and sandstone (Fig. 3) beds that contain very abundant cross strata. Generally, a cycle consists of a lower conglomeratic zone 30 to 100 ft (9 to 30 m) thick overlain by a sandstone zone 20 to 100 ft (6 to 30 m) thick. The conglomeratic zones are composed of beds of poorly sorted pebble to cobble conglomerate and very coarse sandstone. These zones are characterized by lenticular beds 1 to 6 ft (0.3 to 1.8 m) thick, commonly separated from one another by lenses of micaceous sandstone less than 6 in. (15 cm) thick. Internally, the large lenses are horizontally laminated, cross stratified, or both. The cross strata are chiefly trough types and commonly are about 6 in. (15 cm) to 1 ft (30 cm) thick, although sets of cross strata as much as 6 ft (1.8 m) thick have been observed. Preliminary determination of maximum pebble sizes in the conglomerate, excluding the basal conglomerate, indicates a decrease in size southward (Fig. 3).

Measurements of cross strata in the conglomerate revealed southerly or westerly transport directions (Fig. 2), except in the Owlshead Mountains section, where easterly directions were obtained. The directions in the Owlshead may reflect a local western source; and the western edge of the basin (Fig. 2) was drawn with a northerly trend close to this section, based

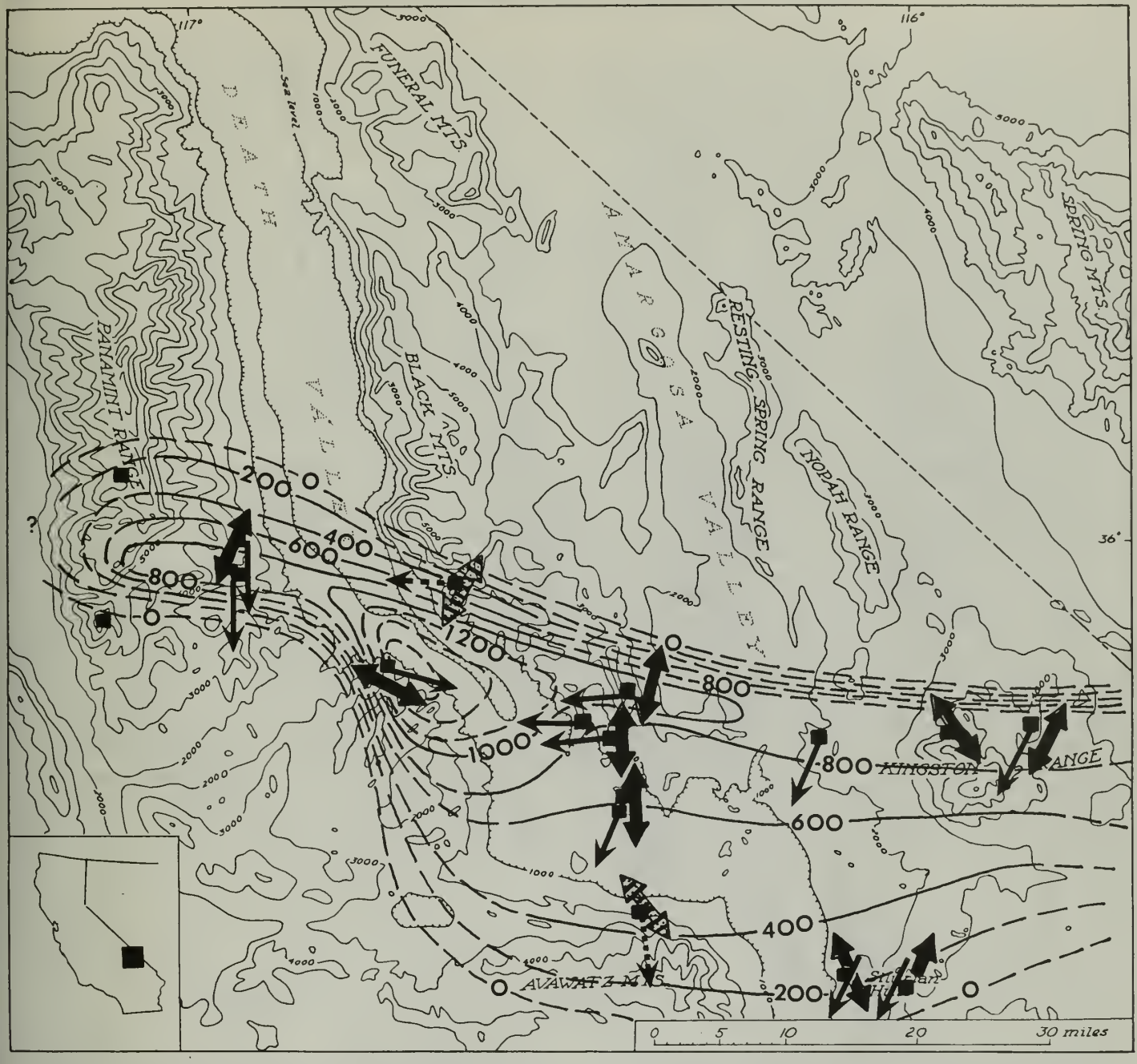

Figure 2. Isopach and paleocurrent map of arkose member. Small arrows are mean directions of foreset dips in conglomeratic units of member. Broad arrows are directions of foreset dips in sandstone units of member. Dashed arrows are probable directions in structurally complex areas. Data represent about 1,200 foreset measurements, 13 measured sections, and observations from additional sections. Isopach interval is 200 ft.

on these data. The cross strata rosettes show scatter, but are unimodal or bimodal with modes generally about 90° apart. The conglomeratic zones are interpreted to be the deposits of braided streams debouching into the basin.

The sandstone zones are composed of fine- to coarse-grained sandstone strata. In some zones, very subordinate, discontinuous beds of chert, shale, and carbonate, a few inches thick, are interbedded with the sandstone. The sandstone is typified by tabular beds that range in thickness from several inches to 3 ft (up to 0.9 m); internally, the beds are horizontally laminated, planar cross stratified, or both. The planar cross strata range in thickness from a few inches to 2 ft (0.6 m). Herringbone cross-strata sets are common in the sandstone zones. Foreset orientations of the cross strata show little scatter, compared to the conglomeratic units, and show unimodal northerly directions in some zones and bimodal northerly and southerly directions in others. The bimodal rosettes have 180° mode separation typical of tidal currents (Fig. 2). The unimodal rosettes that show directions opposite those of the conglomeratic zones represent flood-tide–dominated environments. In the Owlshead Mountains section, the sandstone zones, like the conglomeratic zones, show anomalous directions. The east-west tidal direction of the Owlshead section supports the north-trending border of the basin shown in Figure 2. The sandstone zones are interpreted as deposits of nearshore tidal environments such as tidal deltas.

The cycles are complicated by the presence of zones of interbedded sandstone and conglomerate, which appear either to be due to small-scale interfingering of the fluvial and tidal environments or to be tidal dominated. In addition, the

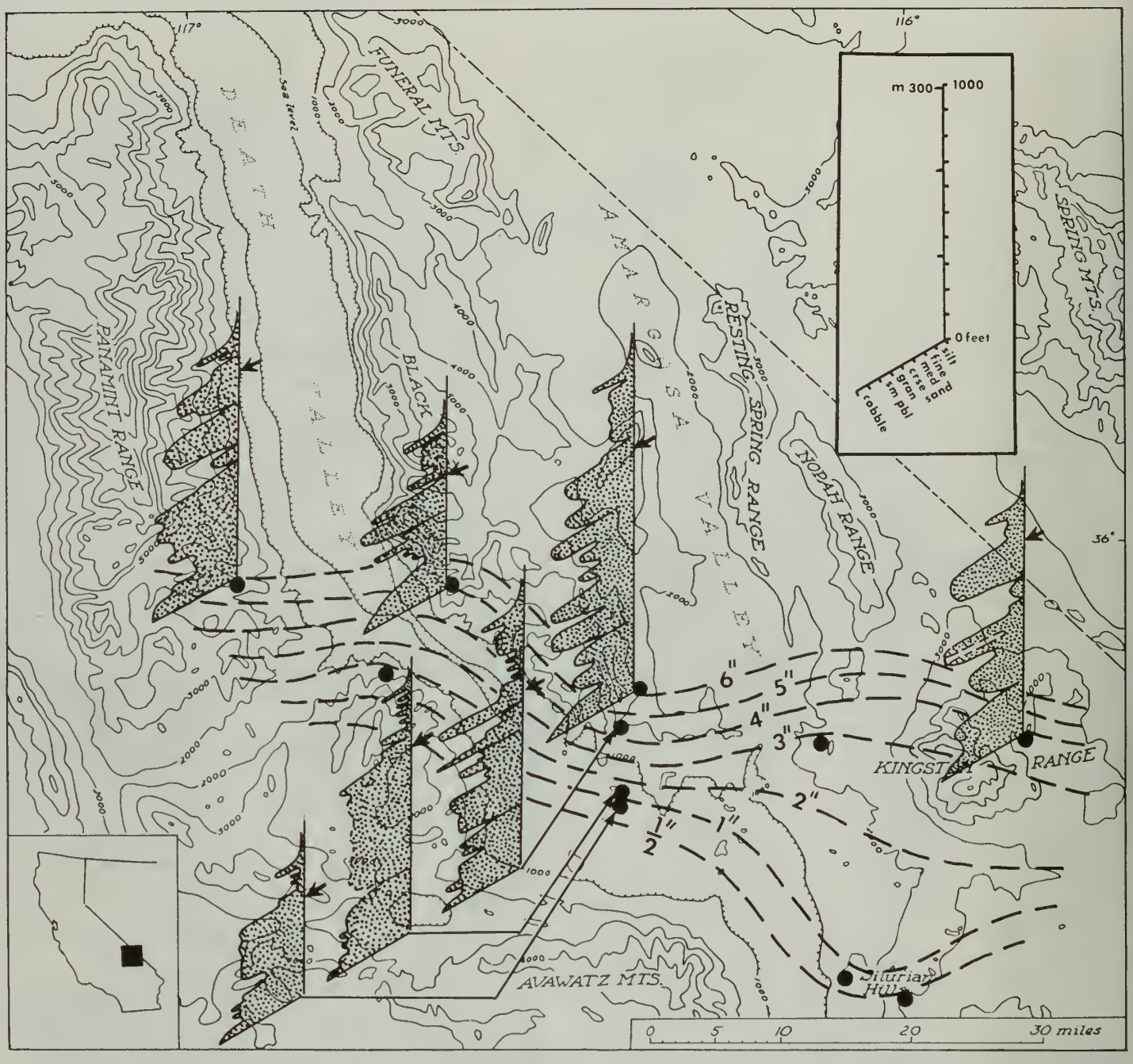

Figure 3. Variations of maximum grain size in lower members. Dashed contours are trend of maximum pebble sizes in arkose member, excluding basal conglomerate. Vertical sections are plots of maximum grain size versus stratigraphic position for selected sections on east-west and north-south lines. Arkose member is from base to top of last conglomerate "pulse" (arrow); feldspathic sandstone member is between arrow and last sandstone "pulse."

sandstone-conglomerate cycles are interrupted high in the member by about 100 ft (30 m) of horizontally laminated sandstone strata. This laminated zone contains scattered, contorted planar cross strata and resembles deposits of modern floods such as those described by McKee and others (1967). On the other hand, this zone may represent a tidal environment with a high regime flow.

The initial site of deposition of the Crystal Spring Formation was a generally east-trending shoreline zone at least 75 mi (121 km) long. Paleocurrent data (Fig. 2) and pebble-size data (Fig. 3) indicate a northern upland source for the sediments. If one applies Sternberg's Law to the pebble-size trends from the central part of the basin (excluding the basal conglomerate), the source is estimated to have been 2 to 4 mi (3.2 to 6.4 km) north of the Eclipse mine area in the central Ibex Hills (point B on inset map of Fig. 1).[2] A close source is also indicated by the angular and arkosic nature of the sediments and the near-

[2] *Sternberg's Law relates pebble size (sometimes stated as weight) to distance of transport. $Y = Y_O e^{-ax}$ where Y is the diameter of the largest pebble a given distance from some reference point; Y_O is the diameter of the largest pebble at the reference point; a was originally stated as a "wear coefficient" but is a constant probably related more to slope factors; and x is the distance between the point at which Y was measured and the reference point.*

Four sections on a nearly north-south line (central panel of Fig. 1) were used for the computation. The constant a was determined using known pebble sizes and distances for this north-south line and the

shore character of the bedforms. Figure 3 also shows that the cyclicity of grain size is best defined in northern exposures and that the "pulses" of conglomerate are more subdued to the south. The cyclicity is interpreted as due to periodic uplift of the northern upland, possibly along a bounding fault zone. The contrast in detectable cyclicity is due to the distance from the bounding zone of the upland.

The bedforms, directional features, and grain-size trends indicate that the sediments were transported generally southward into the basin, where the finer fraction was reworked by generally north-south tidal currents. The cyclicity is attributable to an interfingering of fluvial and tidal environments controlled by periodic uplift in the source area. The member is thus interpreted as deposited in a complex system of nearshore, high-energy environments shifting laterally and vertically through time. Fluvial and tidal sediments are essentially the same in composition and roundness and, therefore, probably resided in their respective environments long enough to acquire distinctive bedforms, but not long enough to become differentially altered before burial.

Abrupt thickness changes within the member probably reflect faulting within the basin during sedimentation, as illustrated in the central panel of Figure 1 and in Figure 6. As the member records high-energy shoreline loci of deposition with a northern upland, finer offshore facies, such as siltstone and shale, might be expected to have been present to the south of the present exposures of the member. However, in later Crystal Spring time (detailed below) the southern region was uplifted and any record of such sediment was probably lost by erosion. An alternate explanation for the lack of this facies would be that little or no fine-grained material was available in the source.

FELDSPATHIC SANDSTONE MEMBER

The feldspathic sandstone member is here defined as the red to purple sandstone, siltstone, and shale that lie above the arkose member and below massive, purple mudstone. The top of the member is everywhere marked by a layer of coarse sandstone, generally only a few feet thick, which is represented by the last "pulse" in Figure 3. The isopach map of the member (Fig. 4) shows similar trends to that of the arkose member, indicating little change in the geometry of the basin. The feldspathic sandstone, however, may have extended farther northward beyond the northern limits of the arkose member.

The basal part of the member consists of tabular beds of fine- to coarse-grained sandstone generally less than 1 ft (0.3 m) thick. Internally, these beds are horizontally laminated, planar cross stratified, or ripple laminated, or they contain some combination of these features. The upper surfaces of the beds are commonly rippled. All types of ripples were observed, including interference ripples, oscillation ripples, lunate/linguoid ripples, and asymmetric ripples. The beds are commonly separated from one another by shaly partings exhibiting mudcracks. The planar cross strata are, in places, arranged in herringbone sets or show evidence of reactivation by opposing currents, as in tidal environments. This basal zone is 100 ft (30 m) or more thick in the central part of the basin, but elsewhere ranges from 30 to 70 ft (9 to 21 m) thick. Orientations of cross strata in the sandstone reveal unimodal northerly current directions at some localities. At other localities, bimodal, generally north-south currents are indicated (Fig. 4). Ripple directions, on the other hand, show multimodal directions typical of intertidal sand bodies. As in the arkose member, the Owlshead Mountains section shows western directions. The current directions are interpreted as due to tidal currents, with the unimodal northerly directions forming in areas dominated by flood tides.

The sandstone zone comprises most of the member at some localities. Ordinarily, however, the zone grades upward to shaly strata which form as much as 60 percent of the member. The sandstone of the lower part of the member generally grades upward into interbedded sandstone, siltstone, and shale which form cycles composed of fining-upward sequences typical of prograding tidal-flat deposition. The sandstone layers contain planar cross strata and ripples like those of the lower part of the member. The silty and shaly layers are thinly laminated and contain abundant mudcracks and shale-chip conglomerates formed from the breakup of mudcrack polygons. These features support an intertidal origin for these rocks. The fining-upward sequences range in thickness from about one to several meters. The thickness of fining-upward tidal sequences approximates the paleotidal range (Klein, 1971); however, the thicker sequences probably reflect subsidence more than the vertical distance between high and low tides. An approximate water depth of 0.5 m in one area was calculated from the grain size and wavelength of oscillation ripples in a fining-upward sequence.[3] The zone composed of interbedded sandstone, siltstone, and shale generally grades upward into shaly strata. Thus, the whole member forms a fining-upward sequence on a large scale, grading upward from sandstone to sandstone, siltstone, and shale and, finally, to shale. Locally, for example, in the Ibex Hills, where the member shows abrupt increases in thickness, it consists of repetitions of the large-scale fining-upward sequence. These thickness changes seem to be related to faulting in the basin during deposition of the sediments, as illustrated in Figures 1 and 6.

The member is interpreted as representing nearshore tidal sand bodies (lower part) and tidal flat (middle and upper part) deposits. The tidal current directions, like those of the arkose member, indicate northerly flood tides and southerly ebb tides. The fining-upward sequences suggest that the tidal flats prograded southward over the basin. The lack of sediment coarser than coarse sand and the abundance, as well as the lowermost occurrence, of shale indicates that the relief of the northern source, which was relatively rugged during deposition of the arkose member, became subdued.

A preliminary study of thin sections obtained from the feldspathic sandstone indicates a greater percentage of quartz than in the arkose member. The quartz content of the sandstone ranges from about 35 percent to about 70 percent. The more quartzose samples contain well-rounded grains and show better sorting than samples from the arkose. The red color of the member is due to abundant hematite in the matrix of the sediments. The relatively high quartz content may be due to removal of feldspar by weathering when the source area was reduced to low relief. It also may reflect a relatively long residence time in the tidal environment. Deeper weathering at the source is

Sternberg equation. In order to determine the distance from the northernmost point (B) to the source of the sediment, only Y_O must be known—that is, the probable maximum pebble size available in the source area. The critical assumption is that the transgressive basal conglomerate contains sizes representative of those available in the source. The maximum size observed in the basal conglomerate was 10 in. (254 mm). The range of 2 to 4 mi (3.2 to 6.4 km) was obtained by using maximum and minimum a values, based on the data, and the assumed 10-in. size for Y_O. Because of these assumptions, the distance to source should be taken as only a "ballpark" estimate. For comparison, even if the source contained boulders 1 m in diameter, it would be only 24 km away.

[3] *Harms (1969) has shown that wave period (T) can be calculated from $T = \pi(E_{max}/U_{w\text{-}max})$ where E_{max} = average ripple crest spacing and U_w (current velocity) must exceed about 19 cm per sec for fine sand. Using crest spacing of 5 cm for one Crystal Spring sample of rippled fine sandstone, T = 0.826 sec. If the maximum depth of wave-affected rippling is about $4/9\lambda$, and $\lambda = (g/2\pi)T^2$, then $4/9\lambda$ (maximum depth for the oscillation ripple above) would be 47 cm.*

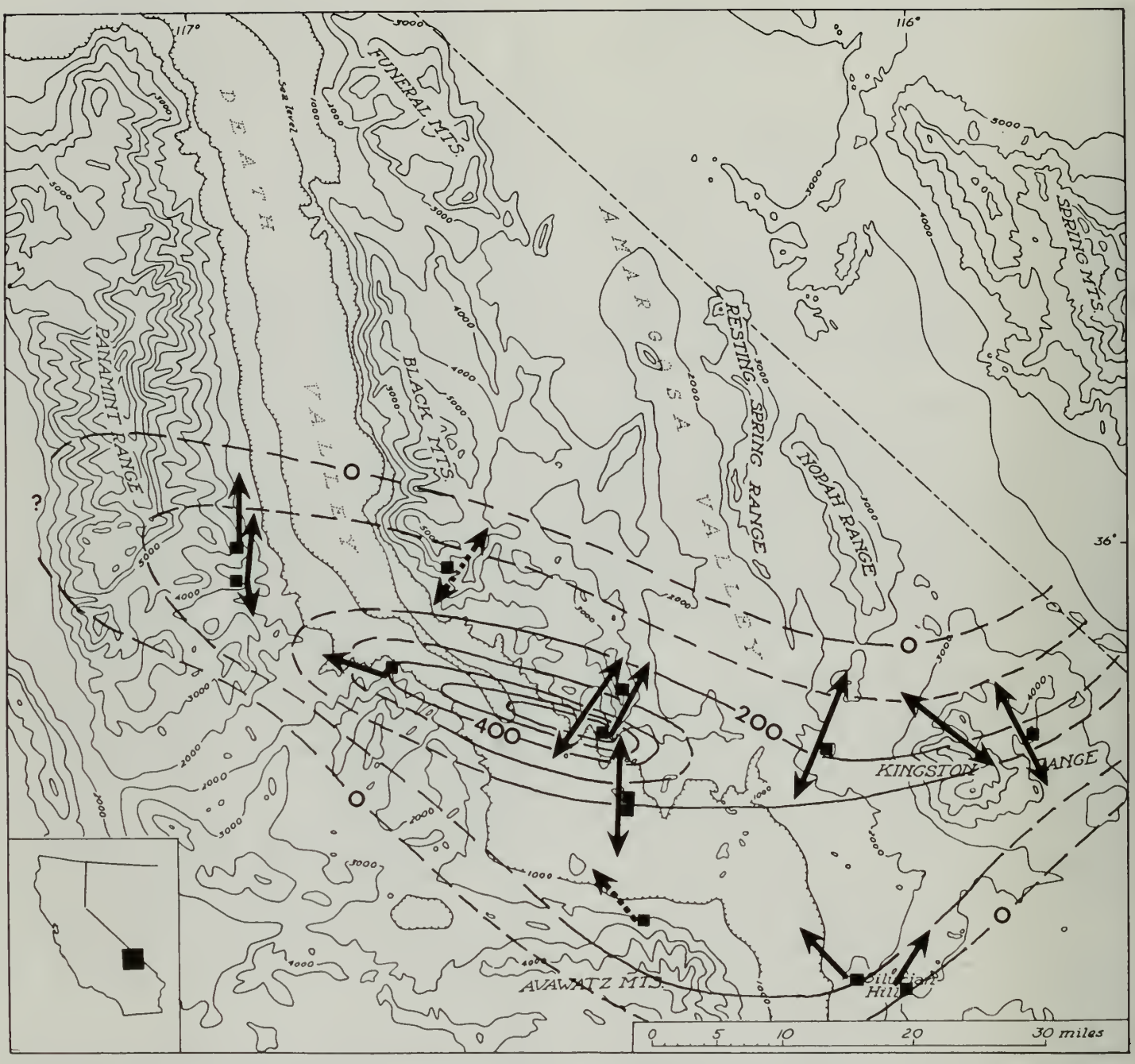

Figure 4. Isopach and paleocurrent map of feldspathic sandstone member. Arrows represent mean directions of foreset dips. Dashed arrows are probable directions in structurally complex areas. Data represent 356 foreset measurements and 13 measured sections. Isopach interval is 100 ft.

indicated by the abundant hematitic material in the matrix of the sediments.

PURPLE MUDSTONE MEMBER

The mudstone member is massive to poorly bedded, purple to red, silty to sandy mudstone that overlies the uppermost sandstone bed of the underlying member. It is overlain by the sandy carbonate zone at the base of the dolomite member. The mudstone member contains scattered grains of sand in its lower part and consists of laminated siltstone and shale in its upper part. The upper part in some areas is metamorphosed below the diabase sill. The member ranges from 30 to 160 ft (9 to 49 m) thick but is commonly about 125 ft (38 m) thick. It is distributed similarly to the lower members, being thickest through the central part of the basin and thinnest in the Silurian Hills. The member is termed a mudstone because it generally lacks regular parting planes, laminations, or bedding; it is poorly sorted and is clayey to silty in composition. A preliminary study of its petrology indicates that 75 to 80 percent of the mudstone is an illite-hematite mixture. The remainder consists of silt and sand grains of quartz (15 to 20 percent) and feldspar (less than 5 percent). The mudstone contains a few thin, discontinuous carbonate beds and very fine sandstone beds and is typified by bluish to greenish reduction spots.

Although the member lies between strata with bedforms that indicate an intertidal environment, it is lacking in distinctive bedforms. The member is apparently the last unit of predominantly clastic material in Crystal Spring time to be derived from the northern upland, although the lack of detailed data from the upper units makes this conclusion tentative. During the deposition of the mudstone, the northern upland was very low in relief. The mudstone may represent low-gradient stream, estuary, or tidal-marsh deposits.

DOLOMITE MEMBER

The dolomite member is here defined as the chiefly dolomitic strata that include the sandy carbonate zone ("upper quartzite member" of Wright, 1968) and the lower part of the "carbonate member" of Wright (1968). The sandy carbonate strata are about 100 ft (30 m) thick across the northern and central parts of the basin and thin southward to less than 50 ft (15 m) in the Silurian Hills. This part of the member does not show a north-south facies change.

In the southern occurrences of the dolomite member, the beds that overlie the sandy zone consist of a zone of dolomite 50 to 100 ft thick (15 to 30 m), in turn overlain by several hundreds of feet of interbedded quartzite, dolomite, shale, and limestone. In the northern exposures, the basal sandy zone is overlain by dolomitic strata as much as 300 ft (91 m) thick. The dolomitic strata display a cyclicity expressed by alternating layers of two types of dolomitic rock. One type consists of light-gray to light-brown dolomite that weathers to various shades of orange. It is very fine grained and contains scattered quartz sand grains, zones of dolomite intraclasts, and subordinate chert beds. It is characterized by wavy to irregular laminations and locally shows ripple cross laminations. The irregular laminae resemble those formed by algal mats. Layers of this type of dolomitic rock range from 1 to 30 ft (0.3 to 9 m) thick and are thickest low in the member (Fig. 5).

The other type of dolomitic rock is very thinly bedded, medium- to dark-gray siliceous dolomite. This type contains thin, wavy laminated beds of dolomite that alternate with thin layers of microcrystalline quartz. On weathered surfaces the latter stand out as brown ribs. The thin layers are typically contorted into folds apparently caused by soft-sediment deformation. The wavy laminae of the dolomite are archlike in places and resemble those produced by algal mats. Layers of this siliceous dolomite range from 1 to 15 ft (0.3 to 4.6 m) thick and increase in thickness and abundance upward (Fig. 5). The top part of the dolomite member in its northerly occurrences consists of a unit of clastic limestone as much as 30 ft (9 m) thick. This unit is composed of sand- to silt-size limestone clasts and contains quartz sand lenses. It is commonly rippled and shows small tidal cross laminations.

Although the dolomite member is about 400 ft (122 m) thick in the Alexander Hills, only 100 ft (30 m) of similar strata can be recognized in the Silurian Hills, 15 mi (24 km) to the south. Either the member simply thins to the south as do lower members, or in the south the upper part is replaced by chiefly quartzitic strata.

The alternation of the two types of dolomitic rock in the northern exposures of the dolomite member is interpreted as due to interfingering of rocks of two depositional environments. The light-gray to light-brown dolomite unit is interpreted as deposited in the higher-energy environment, possibly in a low intertidal or subtidal zone. The siliceous dolomite unit is interpreted as deposited in a lower-energy environment, possibly in a high intertidal or supratidal zone. As the strata that possibly indicate a high intertidal or supratidal environment become more common upward, the sequence can be attributed to prograding (regressive) carbonate tidal-flat deposition. The

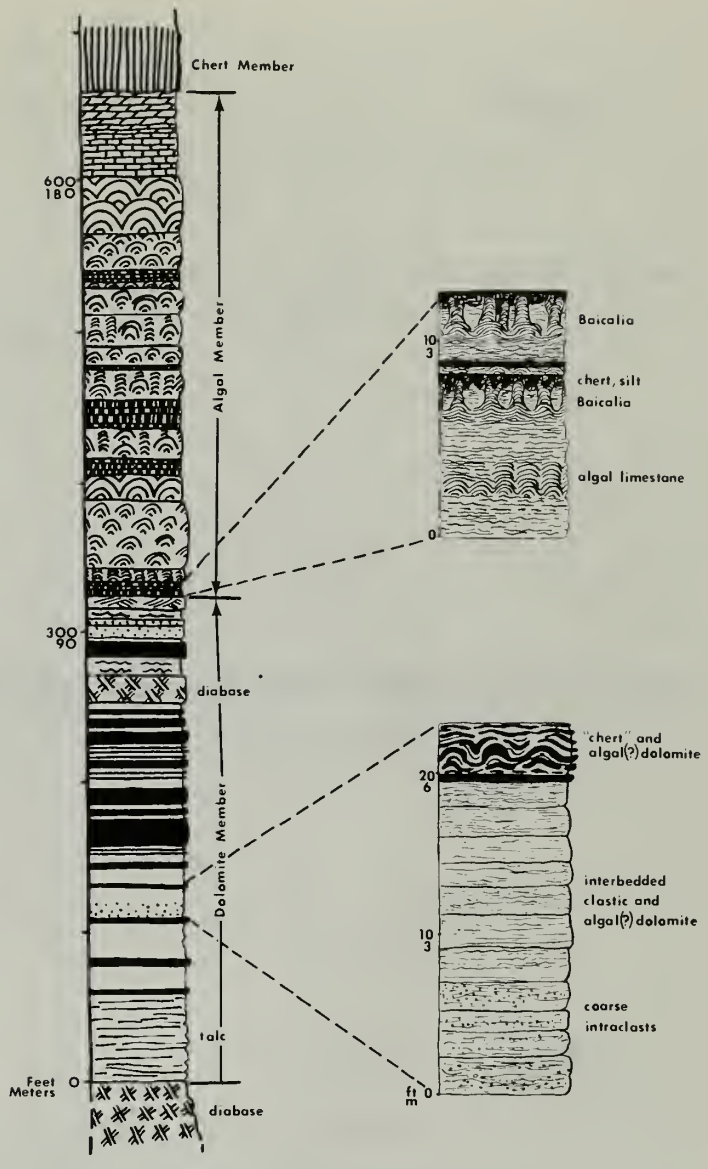

Figure 5. Stratigraphic column of carbonate members in Alexander Hills area. Lower inset details lithologies of single cycle of dolomite member (solid black units of column correspond to upper, ribbed strata of inset; solid white units of column correspond to lower strata of inset). Upper inset details stromatolite cycles within basal zone of larger stromatolitic cycles of algal member. At top of dolomite member is zone of rippled and cross-stratified clastic limestone.

clastic beds at the top of the member in the northern exposures may represent transgression or the return of high-energy tidal environments, possibly related to uplift in the south, as indicated by the quartzose sand in the upper part of the dolomite member there.

ALGAL MEMBER

In the northern exposures of the dolomite member, it is overlain by a carbonate unit, as much as 300 ft (91 m) thick, and characterized by abundant stromatolitic structures. It is here designated as the algal member. It consists of limestone, except in the Saratoga Hills where it is dolomitic. The member contains cyclical repetitions of morphologically different stromatolite forms and includes about 50 ft (15 m) of non-stromatolitic limestone and dolomite at the top. The nature of the cycles varies from place to place. The cycles in the

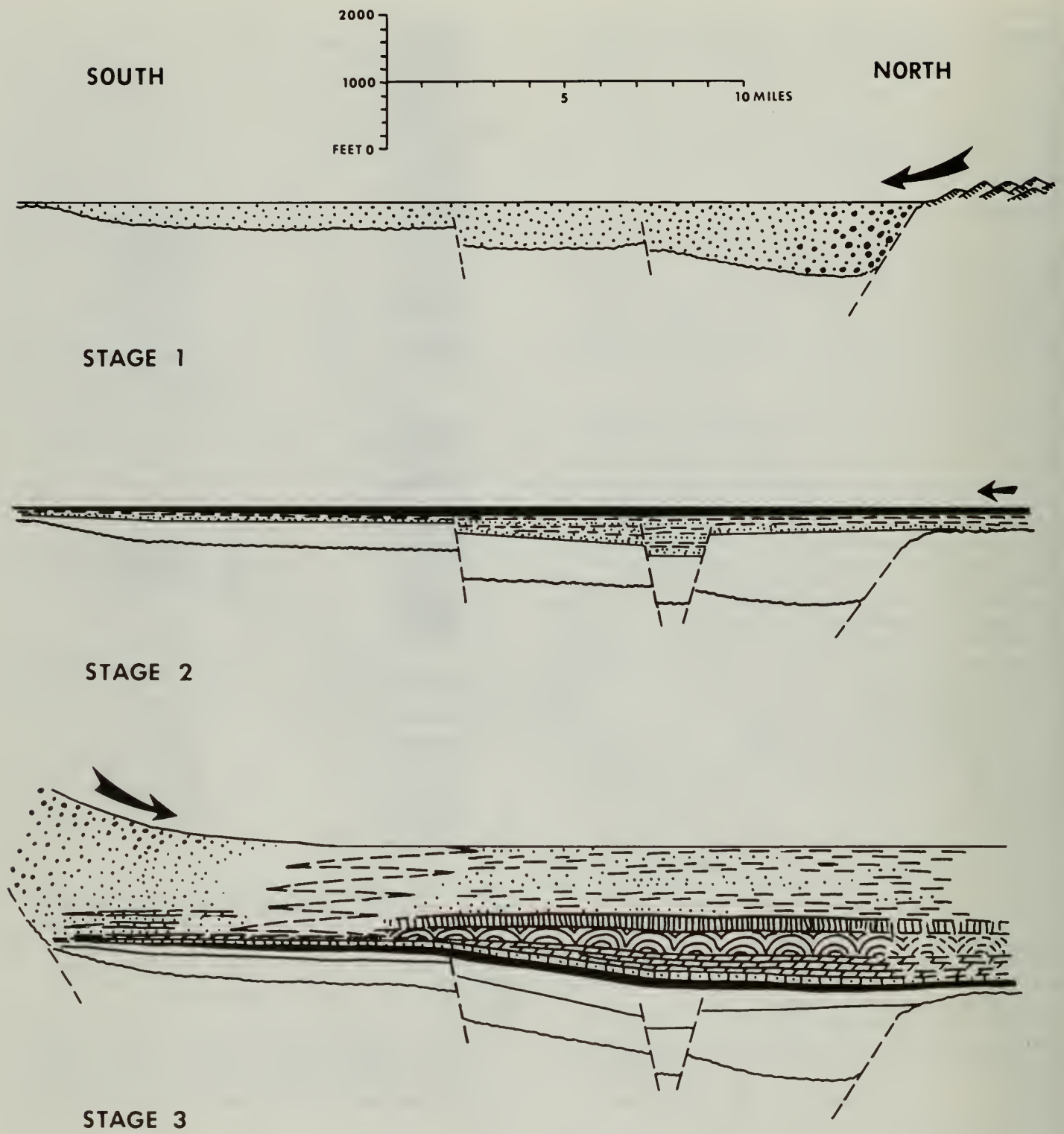

Figure 6. Summary diagram of general development of Crystal Spring basin. Section is composite north-south cross section through center of basin. Exact position of southern upland not known. See text for details of stages.

southern Panamint Range and the Alexander Hills are similar. In the Kingston Range and Saratoga Hills, different sequences of stromatolites were observed.

In the Alexander Hills (Fig. 5), a typical cycle consists of a well-defined basal zone, itself cyclical, overlain by massive zones containing as many as three stromatolite forms. The lowermost cycle in the Alexander Hills is considered to be complete. It consists of a basal zone, 12 ft (3.7 m) thick, composed of cyclical repetitions of the sequence: wavy laminae–algal heads–*Baicalia* columns–silt and chert (Fig. 5, upper inset). This sequence is repeated as many as five times in the basal zone. If intertidal models of Holocene stromatolites apply, this zone may represent transgressive intertidal sequences.

Overlying the basal zone is a bed, 8 ft (2.4 m) thick, consisting entirely of vertically stacked algal heads tending toward columnar development. This is followed by a massive limestone bed, 45 ft (13.7 m) thick, which appears to be composed entirely of algal heads, typically laterally linked, 3 to 4 in. (7 to 10 cm) in diameter. The upper zone of the lowermost cycle is 17 ft (5 m) thick and consists entirely of algal heads about 1 ft (30 cm) in diameter. These upper zones may represent slight regression, followed by relative stability of environment during subsidence prior to the next transgressive basal zone.

Each of the cycles that overlies the lowermost cycle in the Alexander Hills contains the basal *Baicalia*-bearing zone, but the successive zones may differ slightly from those of the lowermost cycle. Commonly the zone characterized by stacked algal heads is missing, and stacked algal heads occur within the zone of laterally linked small heads. The large (1 ft) heads are absent in some cycles. As independent evidence of an intertidal environment for the stromatolites, such as dessiccation features, has not been found, they may have inhabited subtidal areas. The elliptical shape, in plan view, of many of the heads indicates that they grew in current-influenced areas.

In the southern Panamint Range and Saratoga Hills, cylindrical *Conophyton* forms are interbedded with forms possibly of intertidal origin. The relations are not everywhere clear, but *Conophyton* may have grown in low intertidal environments, as in one area they lie between possibly low intertidal and mid-intertidal stromatolites in a vertical sequence. The *Conophyton* cylinders, generally 2 to 3 in. (5 to 8 cm) in diameter and less than 2 ft (0.6 m) long, are inclined and parallel. They apparently grew in that position and may have been aligned with tidal currents.

The algal member of the northern localities probably grades southward into strata composed chiefly of quartzite interlayered with subordinate limestone. The cycles in the algal member thus may be related to periodic uplift of the southern area, which was shedding quartz sand into the basin. The dolomite and algal members record persistent carbonate shelf environments over the northern part of the basin. The northern upland, which was emergent during deposition of the lower members, is not evident in the two carbonate members. The carbonate shelf thus probably extended northward beyond the limits of the lower members.

The large volume of carbonate in these members may have originated from carbonate-secreting algae or from precipitation of carbonate as a result of photosynthetic activity of the algae; however, many of the stromatolite forms are analogous to those created by the trapping of carbonate debris in Holocene stromatolites.

CHERT AND UPPER UNITS

The upper two members of the formation have not been studied in detail. (Their distribution is shown in Fig. 1.) The chert is dark, dense, and fine-grained massive rock, 100 to 500 ft (30 to 152 m) thick (Wright, 1968). The upper sedimentary units are composed of shale, sandstone, and dolomite 200 to 1,200 ft (61 to 366 m) thick (Wright, 1968). Both chert and upper sedimentary units occur only in northern exposures and may be lateral equivalents of quartzite and conglomeratic strata to the south (Fig. 1). These members, then, may be northern basin facies deposited from the rising southern landmass that began to form earlier in Crystal Spring time.

SUMMARY

The Crystal Spring Formation consists of sediments deposited during the initiation of a late Precambrian intracratonic trough. Figure 6 outlines the main stages in the history of the basin in Crystal Spring time. Apparently after an initial marine transgression, a trough formed and was filled with coarse arkosic debris derived from a rugged northern upland that was probably uplifted along a fault (Fig. 6, stage 1). When uplift ceased, the upland was worn down, and, first, tidal-flat sand and mud, then mud possibly of fluvial origin prograded over the basin (Fig. 6, stage 2). The rest of Crystal Spring time was dominated by deposition of quartzose detritus in the southern part of the basin, while, first, algal carbonate shelf sediments, then shale-rich clastic sediments were being deposited over the central and northern parts of the basin (Fig. 6, stage 3). The quartzite and conglomerate of the southern part of the basin were probably derived from a rising southern upland. These major tectonic elements—the trough and the northern and southern uplands—first active in Crystal Spring time, were persistent features, controlling sedimentation throughout the rest of late Precambrian time (Wright and Troxel, 1967; Wright and others, 1974).

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Low-altitude oblique aerial photograph to northeast of west side of southern Panamint Range. Mouth of Goler Canyon is in lower center of photograph. Pole rocks left of Goler Canyon are patch of the platform facies of the Noonday Dolomite. Basin facies crop out south of Goler Canyon (see preceding article). Crystal Spring and Kingston Peak Formations form most of the dark outcrops in foreground (see article by Roberts). Pole rocks (right of center) are Tertiary volcanic rocks. Dark rocks in foreground are intruded by granitic rocks (center and left center). Death Valley is barely discernable beyond the Panamint Range. *Photo 105 by John H. Maxson; courtesy of the National Park Service.*

The Noonday Dolomite and Equivalent Stratigraphic Units, Southern Death Valley Region, California

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INTRODUCTION

The Noonday Dolomite of the Death Valley region of eastern California is generally assigned a Precambrian age because it lies as much as 1,200 m stratigraphically beneath the lowest beds that have yielded Lower Cambrian fossils. It was defined by Hazzard (1937) from exposures near the Noonday mine at the southern end of the Nopah Range. It rests upon various units of the late Precambrian Pahrump Group and also upon a pre-Pahrump complex of crystalline rocks that has yielded radiometric dates as old as 1.7 b.y. Most of the exposures of the Noonday, including those described herein, lie within an elongate area, about 50 km wide, that extends from the southern part of the Panamint Range east-southeastward for about 130 km. The Noonday also has been traced discontinuously northward in the Panamint Range to the vicinity of Tucki Mountain (Hunt and Mabey, 1966), but these more northerly occurrences remain relatively uninvestigated.

In the northern and eastern part of the east-southeast–trending area of exposure, the Noonday consists almost entirely of dolomite. As it contains abundant structures of algal and cryptalgal origin, it qualifies as a shallow-water platform deposit. It passes southward and very abruptly, however, into a deeper water facies of clastic carbonate, breccia, siliceous sandstone, and shale. Because these lithologies are totally unlike those of the Noonday Dolomite as originally defined, this facies warrants and will eventually be given a new formational name. But for the purpose of this report, it will simply be referred to as "the basin facies" and will be divided informally into four members.

In the east-southeast–trending belt of exposure, and for an undetermined distance northward in the Panamint Range, the Noonday Dolomite or "platform facies" comprises two well-defined members. Each has been previously noted and mapped (Wright and Troxel, 1966, 1967; Wright, 1973) but has remained unstudied in detail. These two members will be referred to as the lower and upper dolomite members or as the platform dolomite members. In most of the present exposures of the Noonday Dolomite, the two members display a combined thickness within the range of 60 to 400 m. They terminate southward, generally within a distance of several tens of meters, against units of the basin facies in the manner shown in Figures 2, 3, and 5.

LOWER DOLOMITE MEMBER

The lower algal dolomite is very finely crystalline and laminated. It is pinkish gray to grayish orange on fresh surfaces and weathers to very light gray or grayish orange cliff-forming exposures. The laminations, which commonly are obscured by minute fractures, consist of alternating light and dark layers. Discontinuous accumulations of banded cherty dolomite are common in the upper part of the member. Elsewhere the lower dolomite contains less than 1 percent of insoluble material.

Specimens of the lower dolomite member, as observed in thin section, consist of dark cryptocrystalline dolomite and dolomite spar in varying proportions, and exhibit a spongelike texture. The dark material commonly comprises spherical bodies distributed in a cement of dolomite spar. Variations in the percentages of these two materials account for the laminations.

At many localities, and especially those within 2 km of the platform-basin facies change, the lower dolomite member forms concentrically banded mounds as much as 200 m high and 600 m long (see sketch by Wright and others, 1974b, this volume). These can be interpreted only as algal growth structures and will be described in detail in later publications. The observation that the mound-forming dolomite is identical in physical and chemical characteristics with the more evenly structured occurrences of the member, together with the presence in the latter of smaller scale structures that are apparently of algal origin, has led us to view the entire member as having formed in the presence of algae.

UPPER DOLOMITE MEMBER

The between-the-mound low areas in the paleotopographic depressions outlined by the top of the lower algal dolomite are filled with onlapping younger strata in sharp contact with the underlying dolomite. These compose the lowest part of the upper member and ordinarily consist of dolomite with wavy laminations. Some of the larger low areas, however, are filled with deposits of very evenly laminated rock, composed of interbedded finely crystalline clastic dolomite and dolomitic siltstone, which grades upward into the wavy, laminated dolomite. The clastic dolomite and siltstone are generally dark gray on fresh surfaces, but weather to various shades of red, orange, and yellow. Even where the mounds are absent or only a few feet in relief, such strata commonly directly overlie the lower algal dolomite.

The upper dolomite member, apart from the discontinuous bodies of evenly laminated rock, is generally from 45 to 90 m thick. It is a fine-to-medium crystalline pale gray dolomite, which weathers to light brownish gray and persistently contains from 2 to 10 percent of insoluble residues, consisting of chert, clay, and silt-size quartz. The upper dolomite is less resistant than the underlying member. The lower half consists of internally laminated beds, 2 to 4 cm thick, and commonly displays wavy structures that are generally 10 to 20 cm in amplitude and cuspate to undulating in cross section. Some are clearly cross-laminations and others are slump folds, but most seem best interpreted as laterally linked algal stromatolites.

The upper half of the upper dolomite member consists mostly of interbeds, 20 to 40 cm thick, of dolomite of two types; (1) massive, medium crystalline, and (2) laminated and banded finely crystalline. Distributed through this unit, although more abundant near the top, are moundlike structures with amplitudes of 20 cm to 5 m. Some of these are clearly algal stromatolites of hemispherical form. Other structures are less regular and commonly are surrounded by dolomite clasts. We view these as cryptalgal, yet recognize the possibility that they may be caused by slumping or differential loading.

This mound-forming dolomite high in the upper member is overlain by and intertongued with an intensely cross-laminated dolomitic quartzite unit, which marks a transition into the overlying Johnnie Formation. In the 15 to 30 m of sandstone in which this gradational contact is ordinarily recorded, and in the sandstone that intertongues with the upper dolomite, the quartz grains vary in size from medium to very coarse, and are well rounded, although poorly sorted. The sandstone

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contains both igneous and metamorphic types of quartz; feldspar in proportions of 5 percent or less also is present and consists of albite, microcline, and orthoclase.

BASIN FACIES

The generally southward change from the Noonday Dolomite to the basin facies is so abrupt that the transition is represented in Figure 1 by single lines rather than by gradational symbols. South of these lines the two members of the Noonday are replaced by generally thicker clastic units of the basin facies which generally thicken abruptly. The abruptness of these facies and thickness changes is strongly indicative of vertical movements on fault-bounded blocks. Indeed, through-going faults that were active in Cenozoic and earlier time—the Butte Valley fault in the Panamint Range and the Sheephead fault in the Black Mountains—lie parallel with and close to the lines of facies change. In addition, abrupt variations in the thickness of the platform facies are controlled by vertical movements on contemporaneous faults, some of the faults being detectable in present exposures (Wright and others, 1974b, this volume).

The basin facies, which generally exceeds 300 m in thickness and has a maximum measured thickness of 488 m, is divisable into four members. In upward succession they are here named (1) the basal breccia member, (2) the arkose member, (3) the clastic carbonate member, and (4) the quartzitic clastic dolomite member.

The basal breccia member, which ranges from 3 m to 10 m or more thick in its observed occurrences, is characterized by a polymict breccia featured by gigantic clasts derived from the lower dolomite member of the platform facies and from conglomerate of the Kingston Peak Formation.

Well and evenly laminated dolomite, similar in lithology to the lower dolomite member, commonly forms a thin layer, no more than a meter or two thick, beneath the breccia. It marks a much thinned and basinward extension of the members. Well laminated dolomite also forms masses within the breccia. There the laminations are very irregular and appear to have been deformed contemporaneously with the transportation and deposition of the other components of the breccia. The basal breccia member has been observed only in a northwest-trending belt that lies with 1 km of the southward termination on the zone of marked thinning of the lower dolomite member (Figs. 1 and 5), and is especially well exposed in the southern part of the Ibex Hills (loc. 10, Fig. 1). Elsewhere the arkose member forms the lowest unit of the basin facies. We interpret the breccia as deposited at the base of the platform-basin slope and composed mostly of material dislodged from the slope.

The most complete, easily accessible, and continuous exposures of the other three members of the basin facies occur in the southern part of the Black Mountains immediately north of the highway between Jubilee Pass and the Ashford Mill site (loc. 11 in Fig. 1; Fig. 2), but they also are well exposed at numerous other localities extending from the southwestern part of the Panamint Range south-southeastward to the Silurian Hills.

The arkose member generally consists of arkose and feldspathic siltstone and shale. It forms a north-tapering clastic wedge with a maximum observed thickness of about 250 m. Its disappearance northward ordinarily occurs slightly north of the southern termination of the lower platform dolomite, the arkose thinning as the underlying dolomite thickens. Thus, when traced to its most northerly exposures, it generally can be observed to rest successively upon the breccia and the lower algal dolomite. In most places, however, the arkose member rests upon the Kingston Peak Formation.

In the thickest occurrences of the arkose member, the lower part consists of purple siltstone, shale, and limestone. These pass upward into a sequence of arkose and shale beds that are characterized by graded bedding and by bottom markings that indicate current directions from south to north. The arkose ranges from medium to very coarse grained and is very

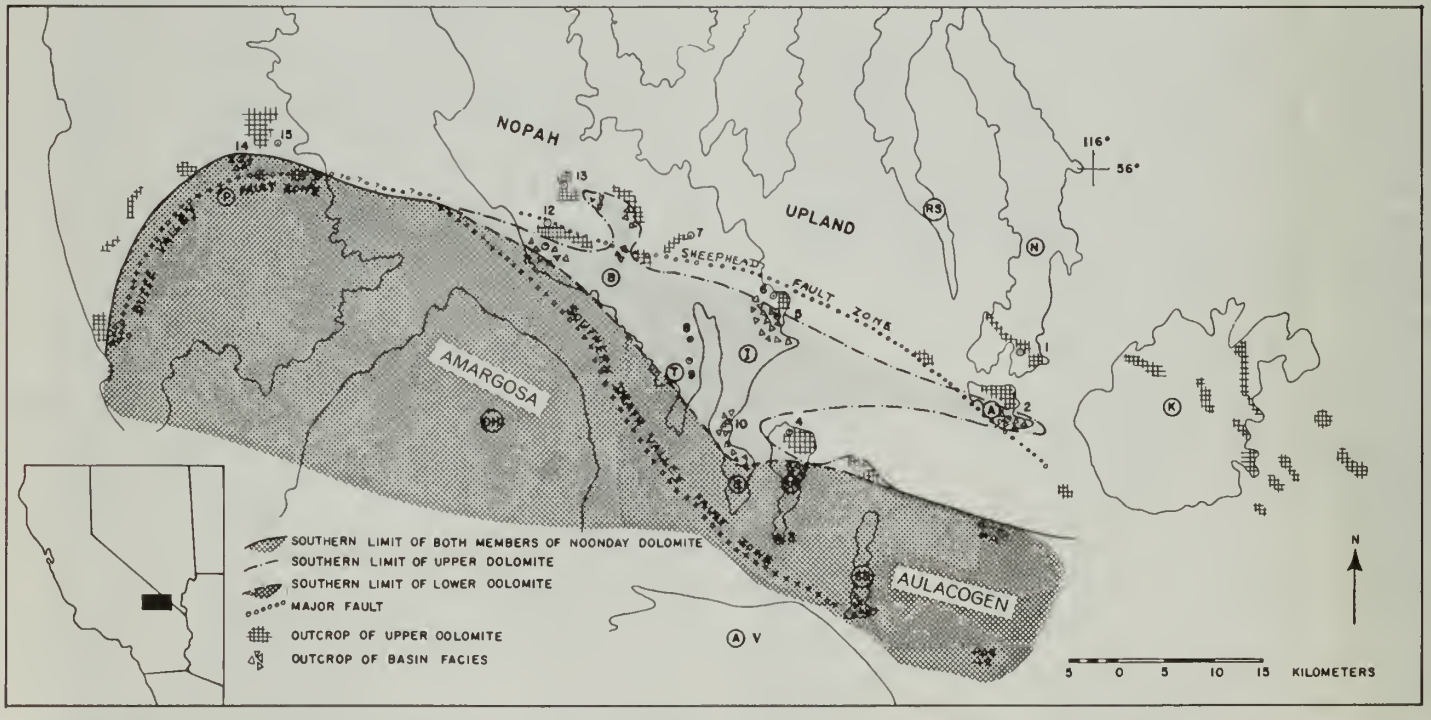

Figure 1 Sketch diagram showing distribution of Noonday Dolomite and equivalent basin facies in southern Death Valley region. Numbers refer to localities where sections were measured. A = Alexander Hills, Av = Avawatz Mts., B = Black Mts., I = Ibex Hills, K = Kingston Range, N = Nopah Range, OH = Owlshead Mts., P = Panamint Range, RS = Resting Spring Range, S = Saratoga Hills, SP = Saddle Pink Hills, SS = Salt Spring Hills, T = Talc Hills.

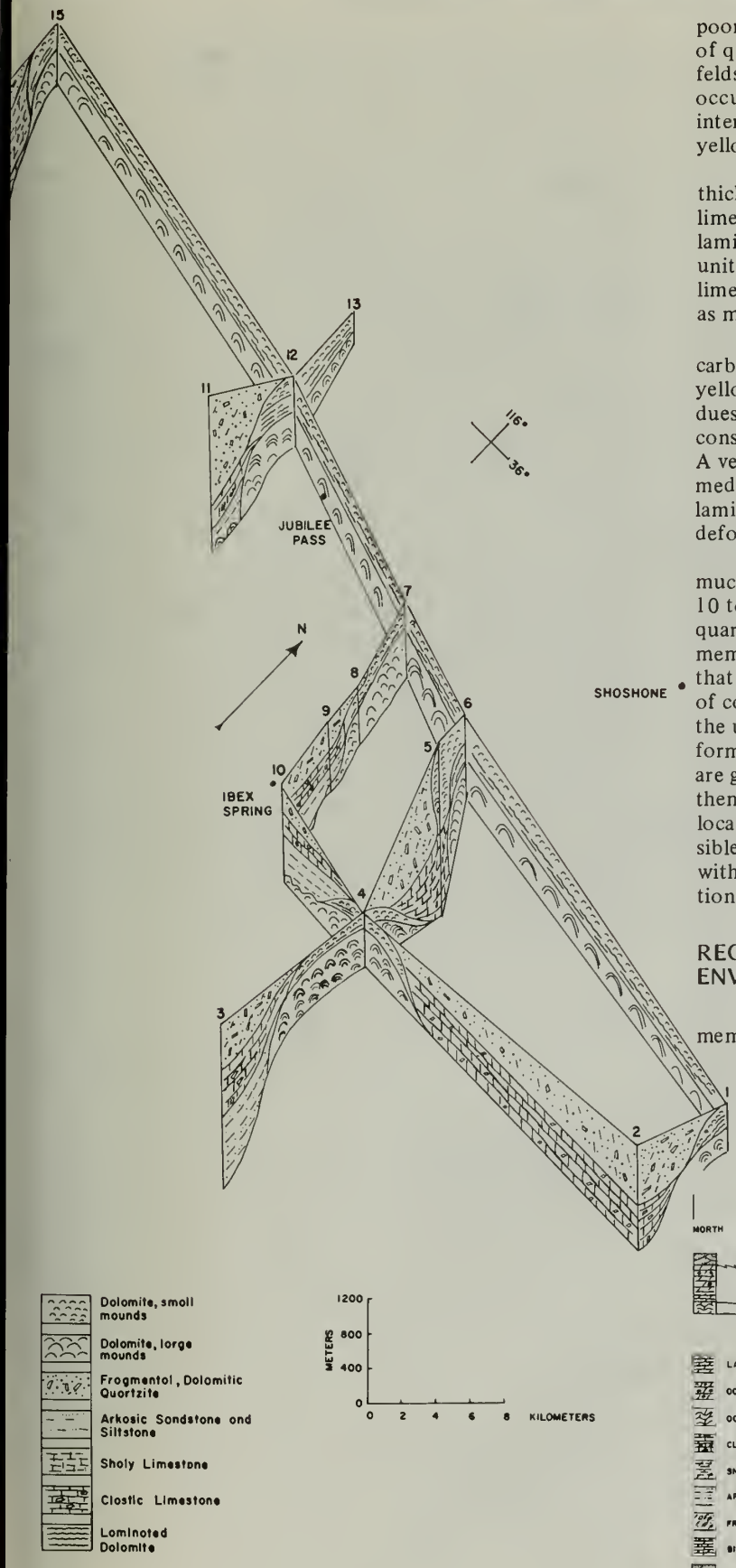

Figure 2. Fence diagram showing relations between members of Noonday Dolomite and equivalent basinal facies. Numbers refer to localities where sections were measured (see Fig. 1).

poorly sorted. It consists of approximately equal proportions of quartz and feldspar, the quartz grains being angular and the feldspar grains highly altered. The thinner, more northerly occurrences of the member consist of purple arkosic siltstone interbedded with argillaceous dolomite beds that weather yellowish brown.

The clastic carbonate member is 190 m in maximum thickness, and consists of varicolored interbeds of dolomite, limestone, and shale that are generally 2 to 30 cm thick and laminated. Close to the platform margin, however, the carbonate unit contains conglomeratic layers with angular fragments of limestone, limy dolomite, and dolomite. Some of the clasts are as much as 15 m long.

Both the bedded carbonate and the clasts of the carbonate-bearing conglomeratic strata weather reddish brown, yellowish brown, and lavender. The content of insoluble residues in the bedded carbonates ranges from 10 to 60 percent and consists of silt-size quartz and chert, illite, and minor kaolinite. A very small proportion of the beds contains well-rounded medium-grained quartz. Some of the beds exhibit contorted laminations that suggest contemporaneous, soft sediment deformation.

The uppermost member of the basin facies, which is as much as 460 m thick, consists of clastic dolomite mixed with 10 to 50 percent of coarse, well-rounded floating grains of quartz and very minor proportions of feldspar. Most of this member qualifies as quartzitic clastic dolomite. Occurrences that lie close to the line of facies change commonly consist of coarse breccia with clasts apparently derived, in part, from the upper platform dolomite. Some of the clasts are intraformational, consisting of the quartzitic clastic dolomite. Beds are generally within the range of 30 cm to 1 m thick. Most of them are devoid of internal structure but cross-bedding is locally present. In many outcrops, bedding is difficult or impossible to detect. The quartzitic clastic dolomite is gradational with the overlying transitional member of the Johnnie Formation, in which cross-bedding is abundant and well preserved.

REGIONAL STRATIGRAPHIC RELATIONS AND ENVIRONMENTS

The regional stratigraphic relations between the two memoers of the Noonday Dolomite and equivalent strata of

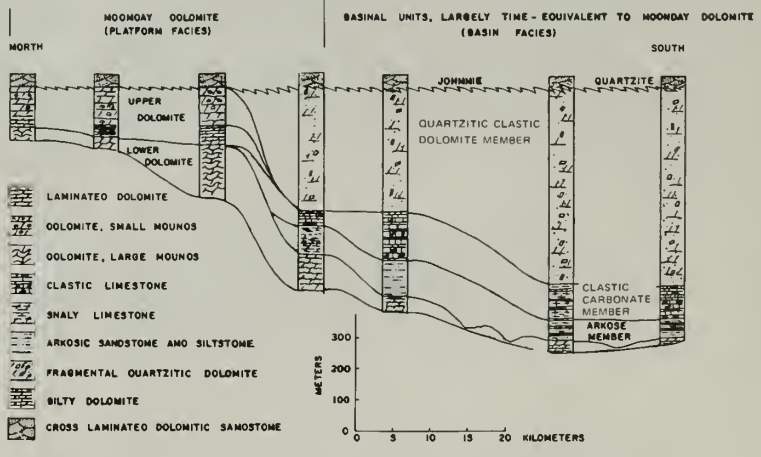

Figure 3. Stratigraphic cross section showing relationships between members of Noonday Dolomite and equivalent basinal facies. Data taken from measured sections near Jubilee mine in west-central part of the Black Mountains and projected to a common north-south line of section. Basal breccia member is missing at this locality.

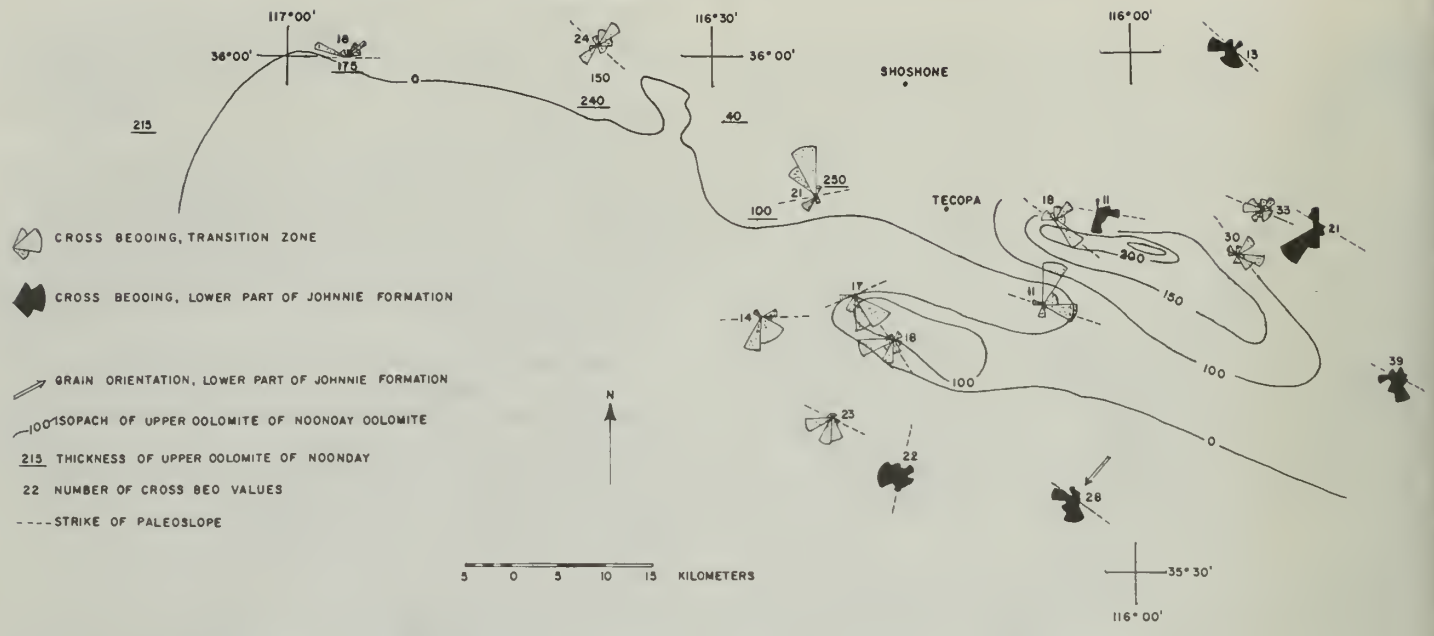

Figure 4. Cross bedding, grain orientation, and thickness information of upper and lower members of Noonday Dolomite, and lower part of Johnnie Formation.

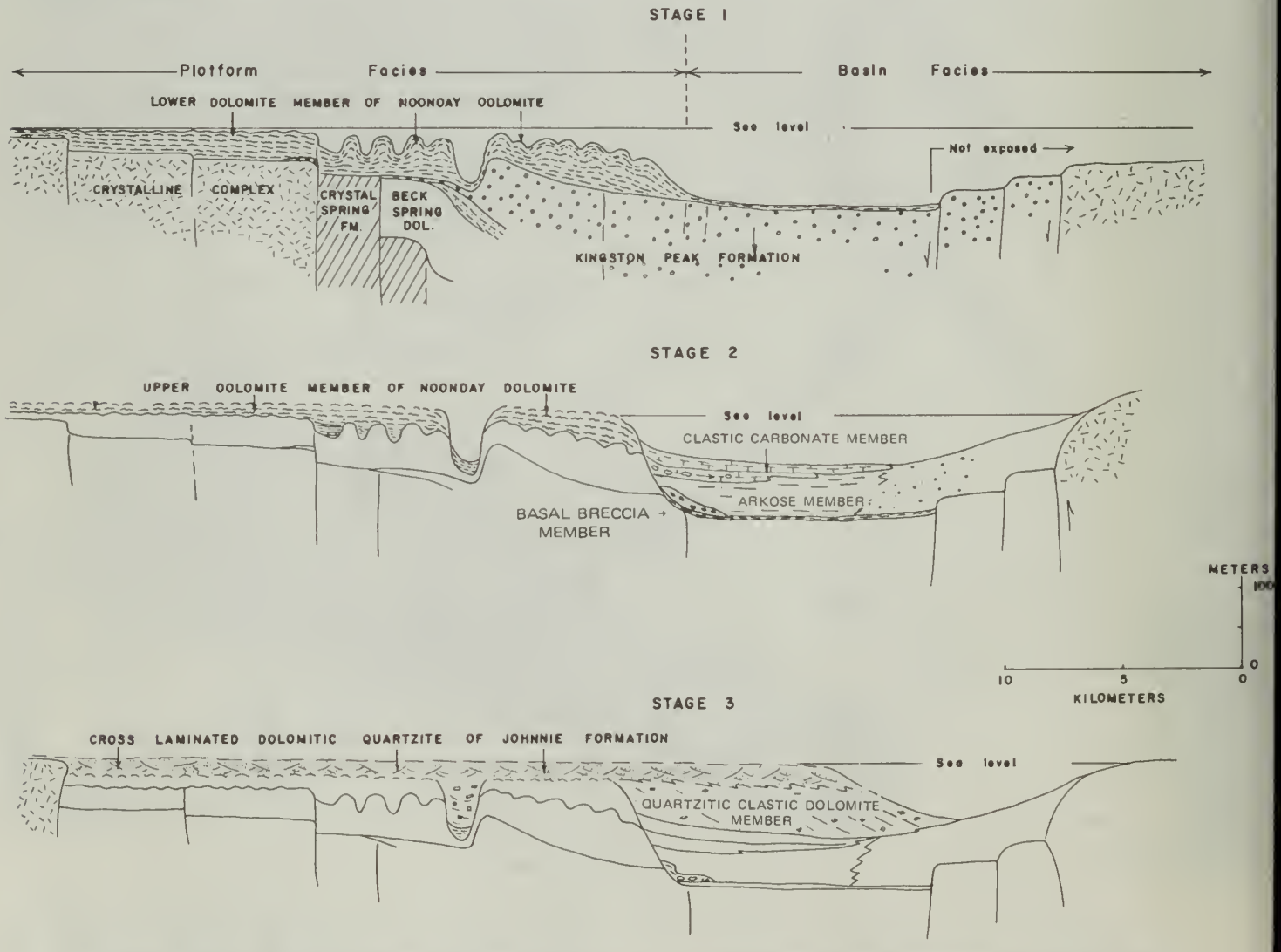

Figure 5. Composite sequential diagram showing development of Noonday Dolomite and equivalent basinal facies in these stages.

the basin facies are illustrated in Figure 3; the boundaries between the two are shown as lines in Figure 1. Observe that the southern limits of the lower and upper dolomite members of the Noonday are divergent in the eastern two-thirds of the region and coincident in the western third. These lines are interpreted as being controlled primarily by vertical movements along faults and contemporaneous with sedimentation. The faults, thus determining the trend of the basin and the adjacent platform.

Verification of the configuration of the northern edge of the basin boundary is independently obtainable through analysis of cross-bedding in the dolomitic quartzite of the overlying transitional member of the Johnnie Formation. A total of 360 measurements were made at the 17 localities shown in Figure 4. The south zero isopach of the upper dolomite marks the platform edge in upper Noonday time. The strike of the paleoslope during the deposition of the lower Johnnie strata, estimated from cross-bedding rosettes, is closely parallel to the trends of the isopachs. The tendency for modal directions in the lower part of the transition member to trend both platformward and basinward probably reflects tidal influences. The generally southwest directions shown by cross-bedding in the higher parts of the transition member may reflect influence of fluvial processes.

The various observations cited thus far are integrated into the composite sequential diagram in Figure 5. In stage one, laminated algal limestone was deposited during transgression of sediment-free marine waters across the beveled fault blocks upon which the Pahrump Group is preserved. Vertical movements on these faults in Pahrump time influenced the thickness and lithologies of the various Pahrump units (Wright and others, this volume). Relief on the surface was generally low, but hills as high as 300 m marked the position of a resistant unit of conglomerate in the upper part of the Kingston Peak Formation. During this first stage of Noonday deposition, the more northerly blocks were depressed slowly, with respect to sea level, and the algal mats grew evenly. We interpret the algal mounds as characterizing blocks that were being depressed less slowly, and that algal growth was significantly favored on highs of the basement topography and retarded on lows. The differential relief of highs and lows was increased by different rates of algal growth as depression proceeded. Blocks that were depressed to still greater depths contain very thin occurrences of the lower dolomite member, or none at all—the growth of the algal mats apparently being inhibited or prevented by insufficient sunlight.

Stage two began with the abrupt ending of the deposition of lower algal strata throughout the platform area, the paleotopography of the surface of that unit being preserved in the sharp contact between it and the overlying upper dolomite. Because this contact is equally sharp throughout the platform area, regardless of the thickness of the lower dolomite member or the presence or absence of the mounds, it seems attributable to a regional event, tectonic or eustatic, that caused a change in sea level. It probably represents a deepening of the ocean floor that killed the mat-forming algae, as no evidence of erosion has been observed even on the highest parts of the largest mounds, and the very evenly laminated silty dolomite and siltstone that fill some of the intermound areas are devoid of algal structures and of features attributable to waves or currents.

The coarse breccia at the base of the basin facies formed after much, if not all, of the lower dolomite member was deposited, as it contains large clasts of that unit as well as much debris from conglomerate of the Kingston Peak Formation, which, immediately north of the basin margin, underlies the lower dolomite member. The breccia probably records a further tectonic deepening of the basin, somewhat before or at the beginning of stage two, steepening the basin margin and favoring the accumulation of the breccia, through submarine sliding, on slopes at the base of the escarpment. The debris from the Kingston Peak could have been derived only from the escarpment, because to the north of the basin margin all older units were sealed beneath the platform cover of dolomite.

The arkose member rests upon basal breccia and lower algal dolomite near the northern margin of the basin and upon the Kingston Peak Formation farther south. Its superjacent relation with the breccia and lower dolomite along the basin margin, however, indicates that some and perhaps most of the arkosic unit was deposited during stage two.

The arkosic sediment provides a record of the elevation and erosion of a granitic land area to the south. Although the granitic terrane has been obliterated by later deformational and igneous events, its southerly location can be inferred from the following observations: (1) bottom markings on arkose beds indicate sediment transport was from south to north; (2) the crystalline basement to the north was sealed by the Noonday carbonate rocks; and (3) the Noonday carbonate and detrital rocks on the shelf contain very little feldspar.

Later, argillaceous limestone and shale were deposited in stage two in the basin, their formation being interrupted by debris flows and turbidity currents that carried broken material from the platform margin and platform-basin slope and deposited it near the base of the slope. The occurrence in the basin deposits of fragments and blocks of limestone, limy dolomite, and dolomite suggests that the algal reef had been partly dolomitized prior to the breaking away of the fragments.

The return, during stage two, of an environment favoring the growth of algae on the platform is recorded in the upper dolomite, but the original limestone of this unit was deposited during an influx of clay and quartz silt. We interpret the wavy structures as forming in the presence of sediment-entrapping algae. The final event, illustrated in stage 3 in Figure 5, is marked by the influx of large volumes of quartz sand bringing the destruction of the reef-forming algal mats. By the beginning of this influx, the reef had been completely dolomitized, because only dolomite fragments are found in the resulting basin deposits of quartzitic clastic dolomite and breccia.

On the shelf, tidal currents reworked quartz and fragmental dolomite into bars and beaches, whereas in the basin, submarine currents carried the quartz and reef debris below the wave base to be deposited as submarine fans. This process continued until the water was sufficiently shallow to permit action of tidal currents. At this point the paleotopography produced by faulting was obliterated. Roberts (this volume) and Diehl (this volume) have presented evidence that the southwesterly paleoslope of the Noonday platform existed as early as the beginning of Pahrump time and continued well after the Noonday Dolomite was deposited. The close parallelism between major faults that have been active in Cenozoic and earlier time and the inferred Precambrian faults that mark the margins of the ancient basin or trough suggest periodic reactivation of ancient zones of weakness over an immense segment of geologic time.

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Stratigraphic Cross Section of Proterozoic Noonday Dolomite, War Eagle Mine Area, Southern Nopah Range, Eastern California

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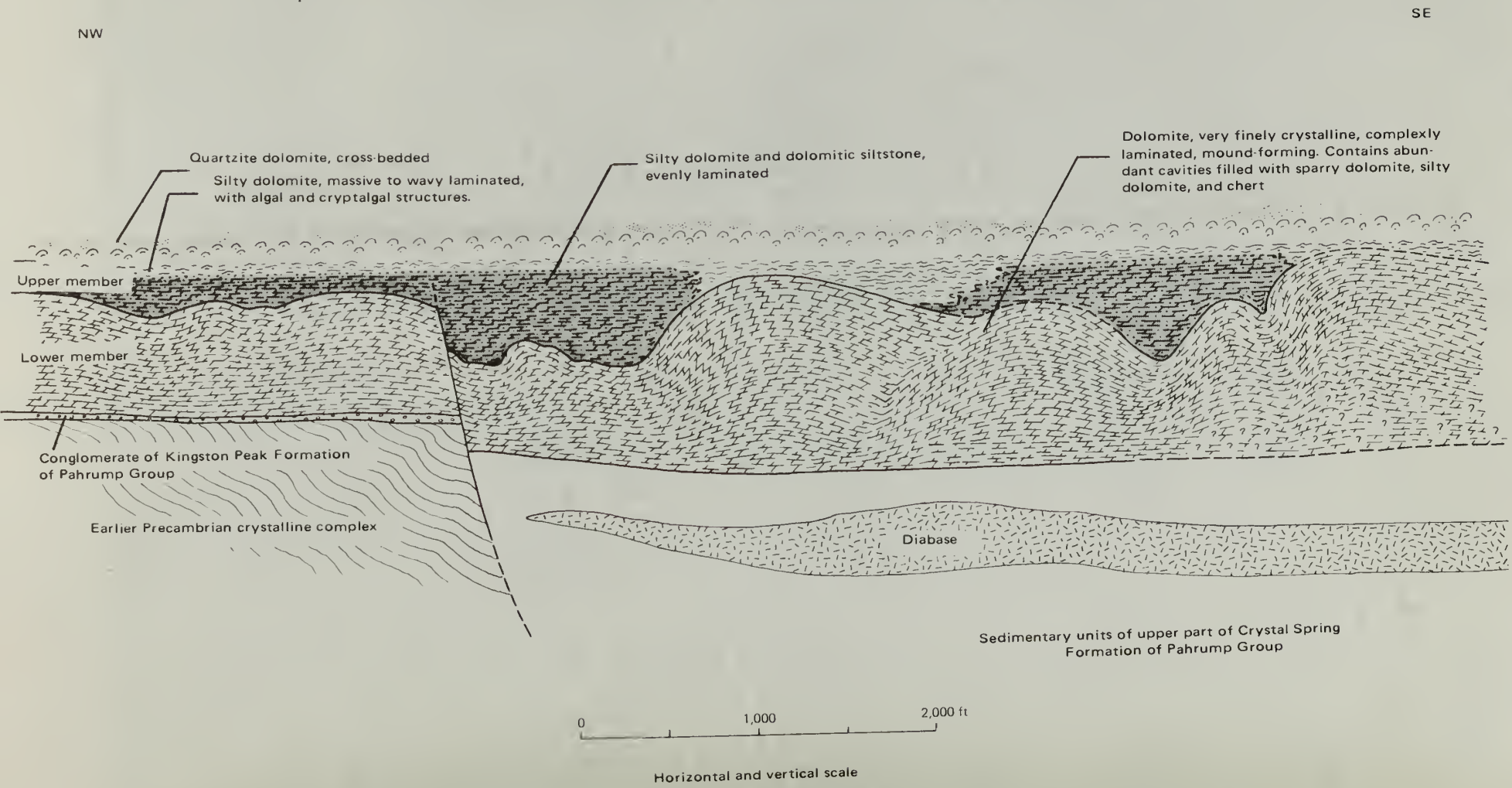
NW
SE
Quartzite dolomite, cross-bedded
Silty dolomite, massive to wavy laminated, with algal and cryptalgal structures.
Silty dolomite and dolomitic siltstone, evenly laminated
Dolomite, very finely crystalline, complexly laminated, mound-forming. Contains abundant cavities filled with sparry dolomite, silty dolomite, and chert
Upper member
Lower member
Conglomerate of Kingston Peak Formation of Pahrump Group
Earlier Precambrian crystalline complex
Diabase
Sedimentary units of upper part of Crystal Spring Formation of Pahrump Group
0
1,000
2,000 ft
Horizontal and vertical scale

Stratigraphy and Sedimentology of the Wood Canyon Formation, Death Valley Area, California

Paul Diehl[1]

INTRODUCTION

Nolan (1924) named the Wood Canyon Formation and described it from exposures in the northwest part of the Spring Mountains, Nevada. Hazzard (1937) extended the use of the name to the southern Nopah Range (Fig. 1), where he measured a section of Precambrian to Cambrian strata including those of the Wood Canyon Formation. Later, Stewart (1966, p. C71) divided the formation into lower, middle, and upper informal members.

Still more recently, Stewart (1970), in a systematic regional stratigraphic study of the Wood Canyon and associated formations of the southern Great Basin, California and Nevada, presented measurements of cross-strata dip directions that he interpreted as indicative of a west- to northwest-dipping paleoslope. The depositional environments of the Wood Canyon, however, remain relatively unstudied.

PURPOSE

Although the general physical stratigraphy of the Wood Canyon Formation has been well established, the sedimentary and tectonic environments in which it was deposited are still little understood, and are the subject of the present investigation. Included in this study are observations of small-scale primary sedimentary features within the formation.

The Wood Canyon Formation is of particular interest because it has diverse lithologies, occupies a stratigraphic position athwart or near the Proterozoic-Paleozoic boundary, and its areal distribution is well documented. The section that contains the Wood Canyon and the underlying Stirling Quartzite, Johnnie Formation, and Noonday Dolomite in the southern part of the Nopah Range has been suspected as a possible stratotype for the Proterozoic-Phanerozoic boundary (Cloud, 1973). In addition, environmental information from this formation, combined with environmental data being gathered by other workers in the area, is useful in attempts to reconstruct the late Precambrian–Cambrian evolution of the Death Valley region.

WOOD CANYON FORMATION

General Features

The Wood Canyon Formation has been divided previously into three informal members that are recognizable over a large area of eastern California and western Nevada. The lower member is composed of interbedded siltstone (36 percent), thinly laminated to platy bedded, fine- to medium-grained feldspathic and micaceous sandstone (50 percent), and laminated, siliceous dolomite (14 percent). The latter occurs primarily as three subunits in the member. The middle member contains a lower sub-unit composed of arkosic conglomerate and grades up-section into cyclically bedded subarkosic sandstone and maroon siltstone. The interbedded light olive-gray to tan siltstone (23 percent) and dark brown to pale red weathering, fine- to very fine grained quartzitic sandstone (49 percent) of the upper member are similar to those in the lower member with respect to distribution and primary sedimentary features. Sandstone beds are 1 to 5 ft thick, well laminated to massive with occasional cross-laminations. The dolomite (12 percent) in the upper member is contained mostly in one sub-unit, which is characteristically cyclically bedded. Distinctive millimeter-sized platelets of echinodermal debris are found in the dolomite beds.

In addition to the gross subdivision of the Wood Canyon into three members, a further subdivision into at least 18 subunits is recognizable. (Detailed descriptions of sub-units recognizable in the southern Nopah Range measured section are shown in Fig. 2).

Lower Member

The lowermost 66 ft of the measured section in the Nopah Range (Fig. 2) are composed of predominantly medium- to fine-grained feldspathic sandstone (76 percent) and siltstone (20 percent) that weathers drab red-brown. In sub-unit SN-3 (Fig. 2), the sandstone content decreases up-section, and sandstone beds form discontinuous wedges and lenses. This decrease in sandstone is accompanied by an increase in siltstone. One of the three dolomite sub-units (SN-2) occurs at 66 ft. The other two are near the top of the member. Sub-unit SN-1 is featured by cyclic bedding, although the scale of Figure 2 is too small to record it. A typical cycle consists of a lower layer of sandstone that is internally massively bedded at its base and contains little mica. It is 1 to 2 ft thick and grades upward into very evenly and thinly laminated, micaceous, very fine grained sandstone. The latter forms the top one-fourth to one-half of each cycle and gives a "shaly look" to the rock that separates the massive beds.

Mudcracks, trace fossils, interference ripples, loading features, scoured surfaces, and discontinuous, convex sandstone bodies that are cross-laminated suggest shallow water deposition. Bimodal and polymodal current rose diagrams are compatible with shallow water marine, probably tidal environments for the lower member.

Middle Member

The basal, ubiquitously cross-laminated arkosic conglomerate (Fig. 2) composes 99 percent of sub-unit SN-10. The overlying subarkosic and feldspathic sandstone and siltstone are cyclically bedded in fining-upward sequences. These cycles range from 1 to 20 ft thick. Most of them are from 3 to 6 ft thick. The base of a typical cycle is marked by a layer of coarse- to very coarse grained quartzitic sandstone or pebbly conglomerate that contains clasts of reddish siltstone. The siltstone clasts apparently were derived from the underlying siltstone bed with which the quartzitic sandstone or conglomerate is in sharp erosional contact (Fig. 3b). The basal layer grades upward into coarse-grained sandstone that is either planar or trough cross-laminated and featured by sets or cosets 2 to 5 in. thick. Planar cross-laminations with straight foresets are as much as 9 in. thick. The next higher layer consists of medium-grained feldspathic sandstone exhibiting complexely festooned cross-laminations and cosets of 2 to 4 in. This sandstone, in turn, grades upward into a massive to well-laminated, fine-grained feldspathic sandstone. Topping the cycle is an alternation of massive and fissile maroon siltstone. The beds within the cycle are 1 to 4 ft thick. The sandstone beds are generally thicker than the siltstone beds, which are as much as 6 in. thick.

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Beds of the middle member are from 1 to 4 ft thick, but change thickness as well as color and lithology laterally. Beds of cross-laminated sandstone commonly grade laterally into siltstone, but also interfinger with siltstone beds. Some of the beds that display the lateral wedging and gradation persist for only 3 to 4 ft, whereas others persist for 100 ft or more.

Upper Member

The upper member is similar to the lower member in thickness and in the proportion of sandstone, siltstone, and dolomite that it contains. The sandstone becomes more quartzitic upward and also shows an upward transition from even-lamination to cross-lamination. The siltstone content increases

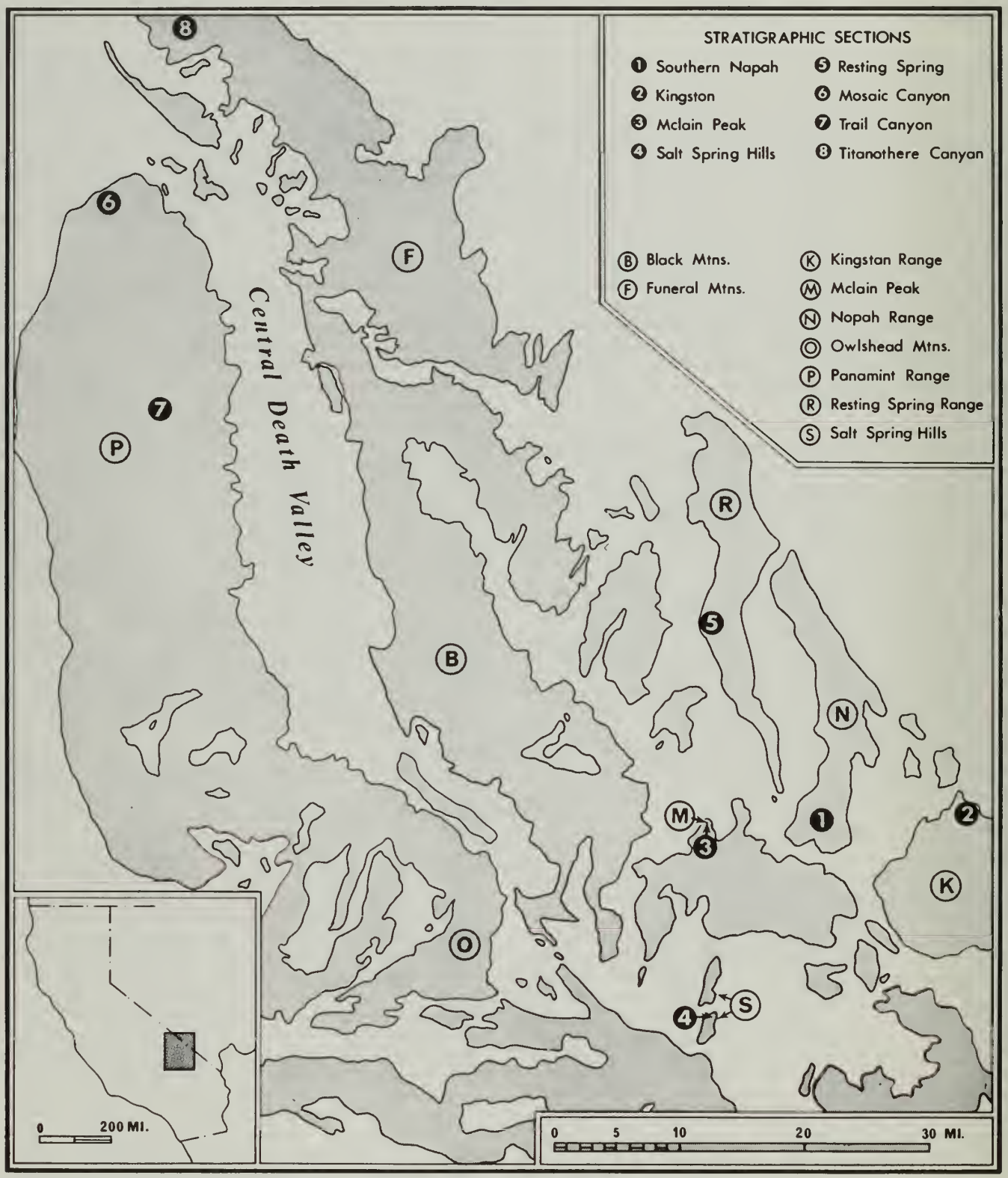
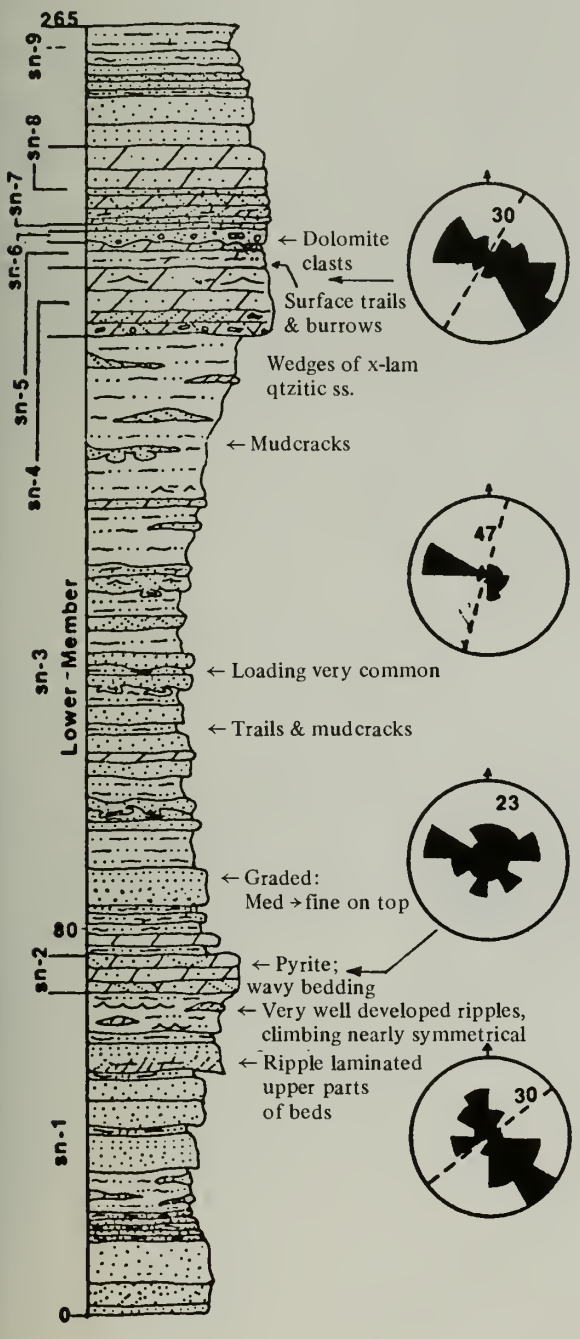


Figure 1. Location of measured sections, Wood Canyon Formation.

up-section, reaching a maximum in sub-unit 18 underlying the dolomite sub-unit. Details of the sub-units, such as first occurrence of fossil hard parts and sand casts of trilobite fragments, are illustrated in Figure 2.

The dolomite sub-unit within the upper member is cyclic in nature. An ideal cycle is illustrated in Figure 3a. This sequence is variable, and one or more lithologies are commonly absent. The cycle, where fully developed, consists of a thin siltstone bed followed vertically by an evenly laminated, well-sorted sandstone that characteristically contains calcareous cement, and locally is faintly cross-laminated. This sandstone is in turn overlain by persistently cross-laminated, dolomitic sandstone in which fossils are common and that grades upward into a cross-laminated to wavy-laminated dolomite displaying abundant quartz laminations and lenses. The sequence is capped by a massive, "elephant hide"-weathering, fine-grained dolomite. Cross-laminations in the dolomite are commonly trough shaped, but planar cross-laminations are also abundant. The sequence is commonly 8 ft thick, and the dolomite beds are as thick as 5 ft. The sandstone and siltstone beds, however, are usually less than 1 ft thick. The siltstone disappears from the sequence upward through the dolomite unit (Fig. 2). The dolomite also contains distinctive platelets of echinodermal debris, which serve to distinguish the upper member dolomite unit from the



sn-9
Intbbb fg, lam ss & qtzt w/ lt. br. to orange tan sltst & slty sh. Few lensatic, 2–6" bdd mod br (5YR3/4) wthrd, lam dol ss; some xlam & ripples. 3' massive vf to fg, pale mod reddish br wthrs pale yelsh br (10YR6/2) in middle. Ss, qtzt & sltst pale red (10R6/2) wthr patchy mod yelsh br (10YR5/4), mica on surfaces. Ss lens and wedge along strike. *Scolithus*?

sn-8
Intbdd dol ss & sandy dol; unit appears striped orange br. becoming mod yel br (10YR5/4) at top. Beds 4 in. at base becoming thicker and more dol up, capped w/ 4 ft vfg med dk gry (n4) dol in 1 ft beds. Ss mg-cg, dk yel br(10YR4/2), xlam in more sandy middle of the unit.

sn-7
Ss, vfg to fg, mod br(5YR4/4) wthr mod br (5YR3/4), 1/16 in. horiz lam, dol peb casts & chips to 2¼ in.; surface trails. Pods of granular ang qtz grains.

sn-6
Granular & vcg, poorly sorted, flat ped cgl ss; chips of dol to ¾ in., pods of gran qtz pebs in 3 in. layers, scoured base. Intbdd fg ss, thin lam qtzt ss w/ ripple lam; some 2 in. beds maroon sltst w/ sand-filled mudcracks or burrows?

sn-5
Intbdd platy, vfg, calc ss to sltst; lt. pinkish to grnsh gry wthrs lt. gry grn to yelsh br. 1–1½ ft beds fg sandy dol, mod br(5YR3/4) wthrs med by (5YR4/4), gritty surface w/ mica, grades lat. to dol ss.

sn-4
Dol, vfg, sandy, med dk gry wthrs dk yelsh br(10YR4/2). Qtz grains raised on dol surface in lower 5 ft. Wavy & xlam wthrs v. dusky br(10R2/2) at base. Pyrite.

sn-3
Intbdd ss, sltst, some slty sh. Ss mica rich, usually platy but beds to 1 ft, mg to fg, tan wthrs dk tan br. Sltst buff tan near base to olive drab w/ red streaks at top. Deformed bdd, loading, xlam, mudcrack, ripples common. Top 35–40 ft slty sh & fissile sltst, ss here 2–6 in. thick lenses of 10–15 ft lat. extent, xlam. Burrows & trails on bdg. Scattered discontinuous dol beds.

sn-2
Dol, mg, med dk gry wthrs lt. med br, sandy lam wthr in relief dusky red br, irregular w/ some xlam. Pyrite rosettes.

sn-1
Intbdd platy, mg to fg, feldspathic ss. Shaly bdd calc ss, v. micaceous. Silica cemented ss beds to 2 ft, med to dk gry wthr drab red br. Lower 10 ft xlam small scale planar & trough foresets. Symmetric ripples well developed in slty top 7 ½ ft. Beds massive base to well lam top. Parting lineations, trails & burrows on bdg surf.

(continued on page following)

Figure 2. Measured section of Wood Canyon Formation, Nopah Range. Section measured on southern slope of hill VABM 4238, approximately 1½ miles north of the Noonday mine, sec. 11, T20N, R8E, Tecopa, Calif. 15-minute quadrangle. Dashed lines on current rose diagrams indicate possible strike of paleoslope. Scale 1 in. = 40 feet.

other dolomite beds both within the Wood Canyon Formation and in the overlying and underlying formations.

TRENDS IN THE WOOD CANYON FORMATION

Cross-lamination measurements were recorded with close reference to their stratigraphic positions so that vertical variations in their orientation, discussed below, could be detected within each stratigraphic section. The lateral variation of preferred orientation of paleocurrent indicators could be investigated by comparing their trends from section to section within the Death Valley area.

Vertical Trends

Analysis of the orientation data gathered so far reveals a consistent vertical variation of paleocurrent orientation within each stratigraphic section. This variation can be correlated with the succession of sub-units defined above (Fig. 2). The bimodal and polymodal frequency diagrams of cross-strata measurements taken in the lower member are consistent in each section measured (Fig. 4a) and suggest a northwest-southeast paleocurrent orientation.

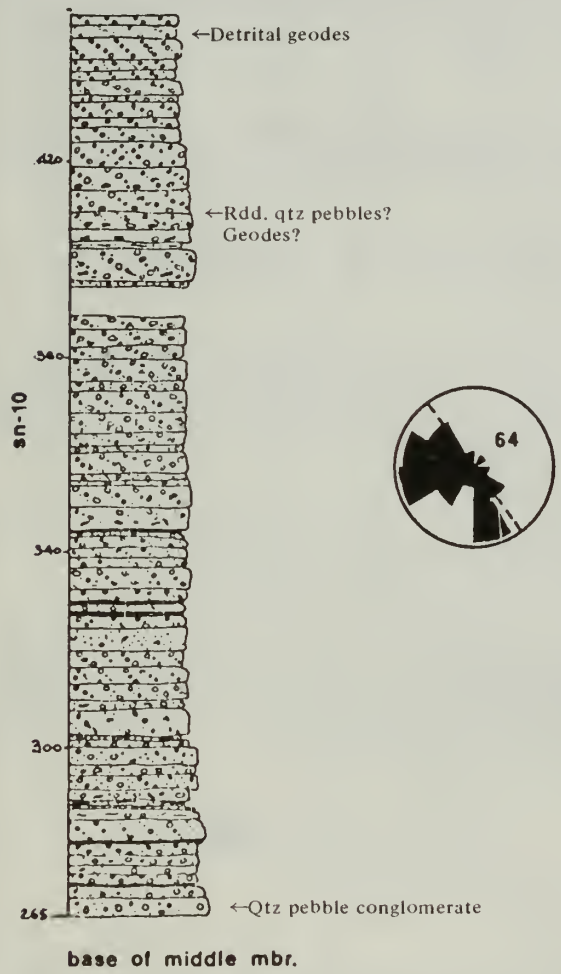
In marked contrast to the lower member, the paleocurrent indicated by current rose diagrams taken from the arkosic conglomerate at the base of the middle member, is to the southwest (Fig. 2, SN-10; Fig. 5a). This nearly 90° change in paleocurrent occurs abruptly at the contact between the lower and middle members. Up-section in the middle member, cross-lamination orientations show a re-establishment of a northwest-southeast paleocurrent. This change is gradual; the frequency modes of paleocurrent direction measurements from sub-units in the ramainder of the middle member show an up-section clockwise rotation.

Polymodal and bimodal cross-lamination frequency diagrams plotted from data taken in the upper member again reflect a northwest-southeast paleocurrent as defined in the lower member.

Regional Trends

Sub-units recognized within the southern Nopah Range can be recognized in sections throughout the Death Valley area, particularly if sub-units five, six, and seven of the lower member are grouped into one, thus leaving seven regionally recognizable sub-units of that member. In general, these sub-units occur with little lithologic variation throughout the Death Valley area, so that distinct lateral facies are not generally recognizable. The current-direction indicators that are measured from correlative sub-units in various stratigraphic sections (Fig. 1) yield consistently oriented paleoslope indicators except where tectonically disturbed (Figs. 4a, 5a). The same vertical pattern of paleoslope variation described above can generally be detected in sections thus far measured.

Although the nearly 90° rotation of current direction indicators up-section is clearly demonstrated in the Mclain Peaks section, the rose diagrams there are anomalous, as they suggest a northward paleoslope for the lower and upper members and a northwest paleoslope for the arkosic conglomerate at the base of the middle member. The differences between these orientations and the others already described seem best explained



sn-10

First 4′ granular to vcg qtz peb cgl, poorly sorted vitreous white & transparent ang qtz fragments; contains specks of dk yelsh orange(10YR6/6) and lt. grn clay material, bed wthrs pale grn(10G6/2). Basal ctc sharp and uneven but top transitional w/ overlying dk grnsh gry(5G4/1) arkosic cgl, vcg & granular, planar xlam v. abundant w/ some trough foresets. Rare grysh-red purple (5RP4/2) sltst in beds 6″ or less. Felds abundance patchy. Beds 2–5′ w/ xlam sets up to 1′.

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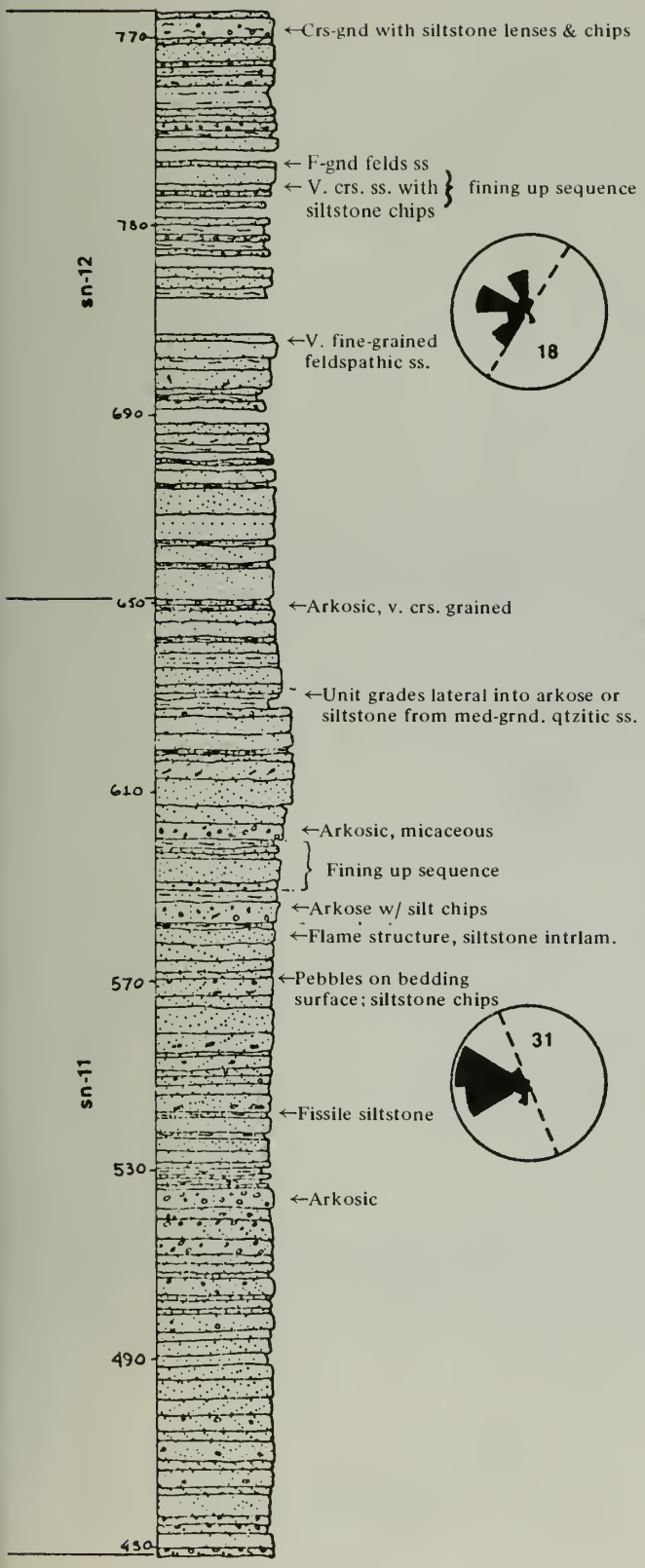
Figure 2 (continued).



sn-12

Cyclic intbdd ss & sltst. Ss are (1) fg, felds, grysh red purple (5RP4/2) wthr grysh red(5R4/2), beds 15″–4′, more abundant & thicker upward, horiz lam to massive w/ few planar xlam; (2) m–vcg ss, pale red purple(5RP6/2) wthrs streaky pale pink & grysh red, complex trough xlam, ang & rdd sltst flat peb chips to 2″ abundant, not well cemented, beds 1′ & less; (3) well cemented, m–cg, grysh purple(5P4/2) w/ some white to v. lt gry(N9–N8) wthrs mod yelsh br(10YR5/4), ang to subrdd qtztic ss, well defined complex trough xlam w/ long tangent foresets, also planar straight foresets truncated top & bottom, cosets 2–3″ but up to 1′ when planar, beds 2–4′ pinch & swell w/ lithologies grading laterally. Fissile sltst, much mica on bdg planes, beds to 2′, dusky red(5R3/4) to grysh red purple(5RP4/2).

sn-11

Intbdd qtz peb cgl, arkose cgl, qtzt & felds ss. Felds ss pale red(10R6/2) to grysh red(10R4/2), mg, very well developed complex trough xlam w/ cosets 2–8″. Planar xlam w/ larger sets up to 1′. Qtz peb cgl w/ pebs of qtz, red chert, & detrital geodes, beds med yelsh br(10YR5/4) & lt gry(N7). Arkose cgl becoming less frequent. Sltst becomes more abundant than in SN-10 w/ some beds to 1′. Lithologies form fining upward sequences of 7–14′ thick. Beds not laterally continuous, pinch & swell, wedge out, & grade laterally

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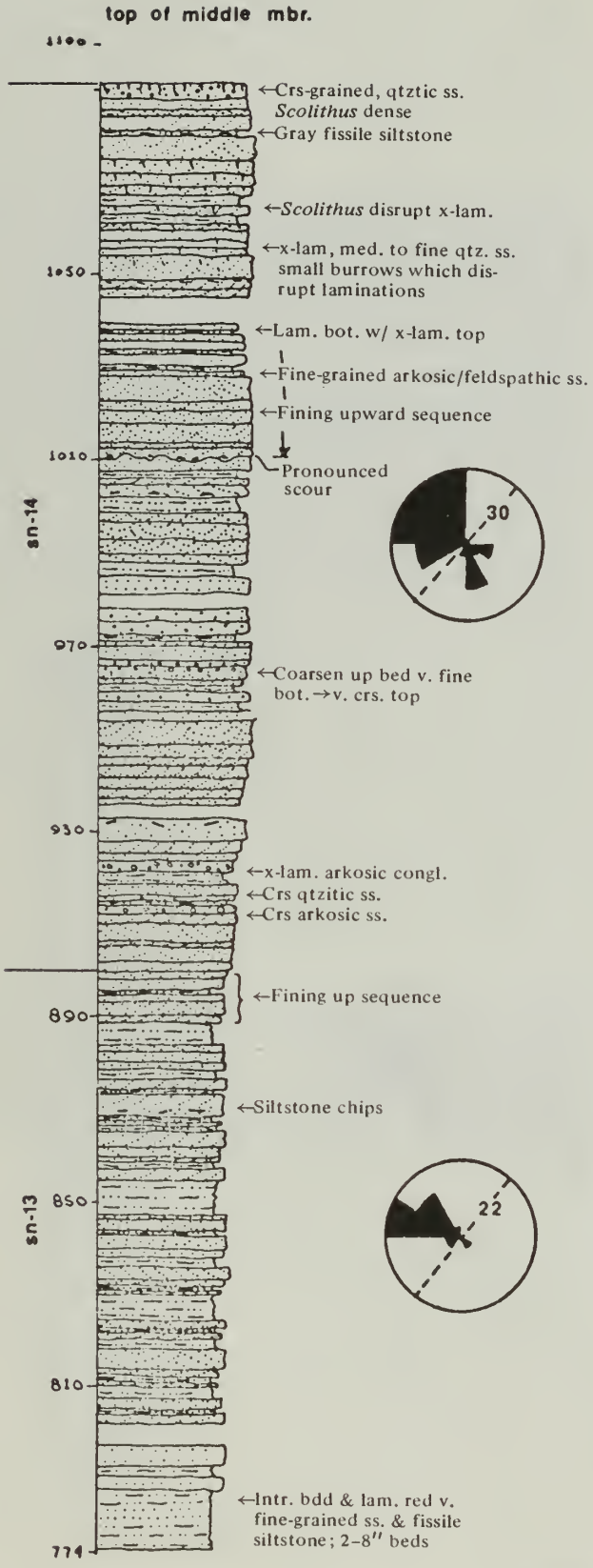
Figure 2 (continued).



sn-14

Qtztic & gwke ss, well defined complex xlam, troughs & complex v. tangential foresets, beds 2–3′ w/ cosets 3″ to 1′ but 3–8″ most common, m–vcg w/ cg predominating. Vcg & gran arkosic beds comprise 12′ of section 52′ from base of unit. Top of unit cg to vcg, well cemented qtzt ss w/ v. densely spaced *Scolithus*. Fissile sltst & massive fg felds ss rare.

sn-13

Dominated by fissile maroon sltst w/ beds up to 3′ thick, and grysh red purple felds to gwke ss, mg w/ parts of section composed of as much as 12′ continuous ss in 1–2′ beds. Chips of sltst v. common on bdg surface and w/in beds. Other lithologies of SN-11 & 12 also occur. Fining upward sequences.

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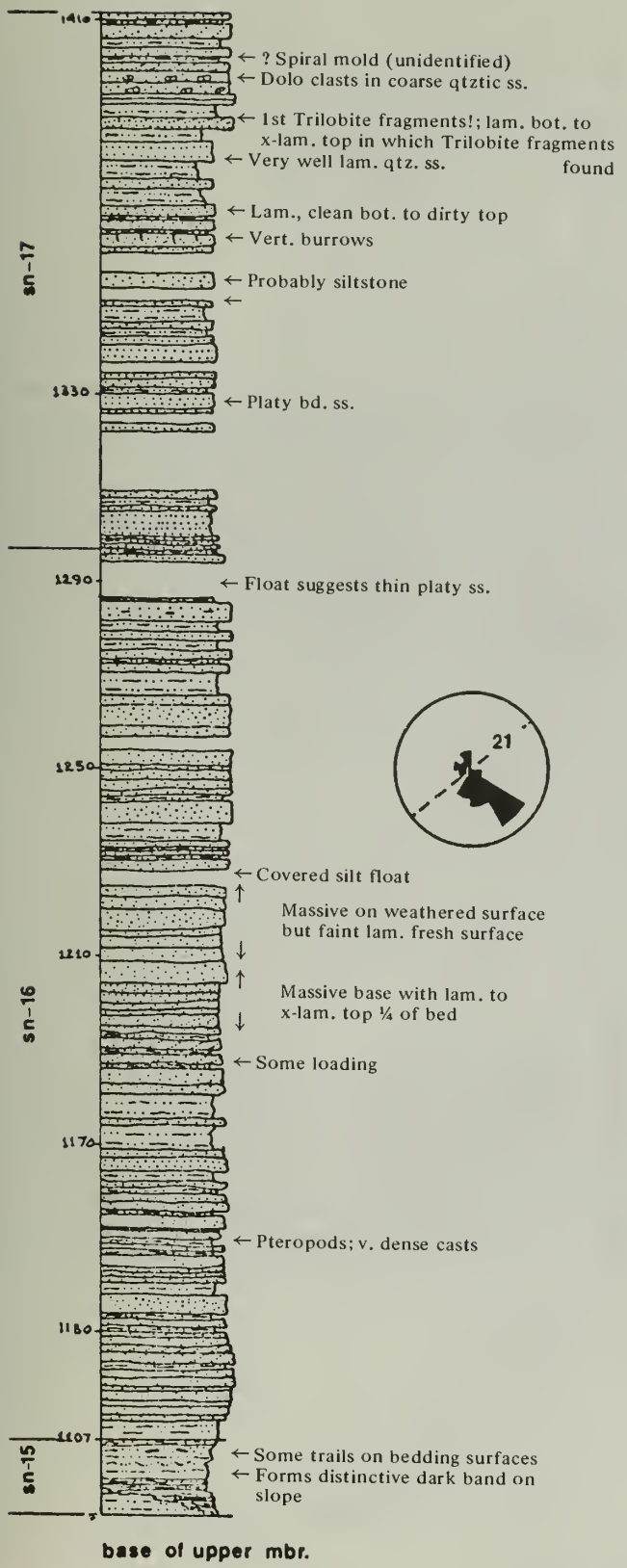
Figure 2 (continued).


1410
← ? Spiral mold (unidentified)
← Dolo clasts in coarse qtztic ss.
← 1st Trilobite fragments!; lam. bot. to x-lam. top in which Trilobite fragments found
← Very well lam. qtz. ss.
← Lam., clean bot. to dirty top
← Vert. burrows
sn-17
← Probably siltstone
1230
← Platy bd. ss.
1290
← Float suggests thin platy ss.
21
1250
← Covered silt float
Massive on weathered surface but faint lam. fresh surface
1210
Massive base with lam. to x-lam. top ¼ of bed
sn-16
← Some loading
1170
← Pteropods; v. dense casts
1180
1107
sn-15
← Some trails on bedding surfaces
← Forms distinctive dark band on slope
base of upper mbr.

sn-17

Qtztic ss, fg & vfg, well cemented, thin horiz lam, ss dk grnsh gry(5G4/1) wthrs vary colored dk yelsh br(10YR4/2) & dusky yelsh br(10YR2/2). Covered section may be vfg platy bdd ss & sandy grysh grn sltst–seen in float. Beds 9″ to 5′ but most 1–2′. Some massive & platy bdd "sugar-like" qtzt, lt gry(N7) wthrs mod orange pink(5YR8/4) & yelsh gry(5Y7/2). Xlam ss 72′ from base & becomes more abundant up-section. Top of unit has fg qtztic ss w/ lens of vcg qtz & flat rdd dol cobbles 1″ to 3½″ long, xlam. First trilobite casts occur in place 92′ from base of unit. Trilobite frags. associated w/ lt to med gry(N7–N5) qtztic ss, horiz lam base & xlam top. *Scolithus*, bioturb. lam also found. Unidentified spiral mold found in xlam calc cemented qtztic ss at top of unit.

sn-16

Slope forming intbdd vfg–fg ss & sltst (60/40). Sltsts dk yelsh br (10YR4/2) shaly & lt olive gry(5Y5/2) massive & fissile. Sltst beds sandy & grade transitionally w/ ss. Ss are pale yelsh br (10YR6/2) to med dk gry (N4) to olive gry(5GY4/1) w/ some pale red wthrng to dk redsh br(10R3/4), grysh red(10R4/2) & dusky yelsh br(10YR2/2), but distinctive pinkish or redsh br wthrd surface. Beds massive, 9″ to 2 1/2′. Lower 8–12′ of unit gwke ss becoming fg & qtztic up w/ decrease in shaly sltst. Xlam rare in lower 1/4 of unit, some scour w/ sltst chips. Pteropod bed 3″ thick 9′ from base. Trilobite frags. in float in upper part of unit.

sn-15

Gwke ss, m–fg, grysh br wthrs blksh red(5R2/2), poorly cemented, lam w/ thin shaly wavy lam approaching xlam, platy bdd but some beds to 15″. Cg patches, poorly sorted, ang qtz grains & blk sltst/sh chips. Intercalated dk maroon sltst. Upper ctc transitional. Forms distinctive dk wthrng band. 6′ olive gry(5Y4/1) sltst occurs NW along strike.

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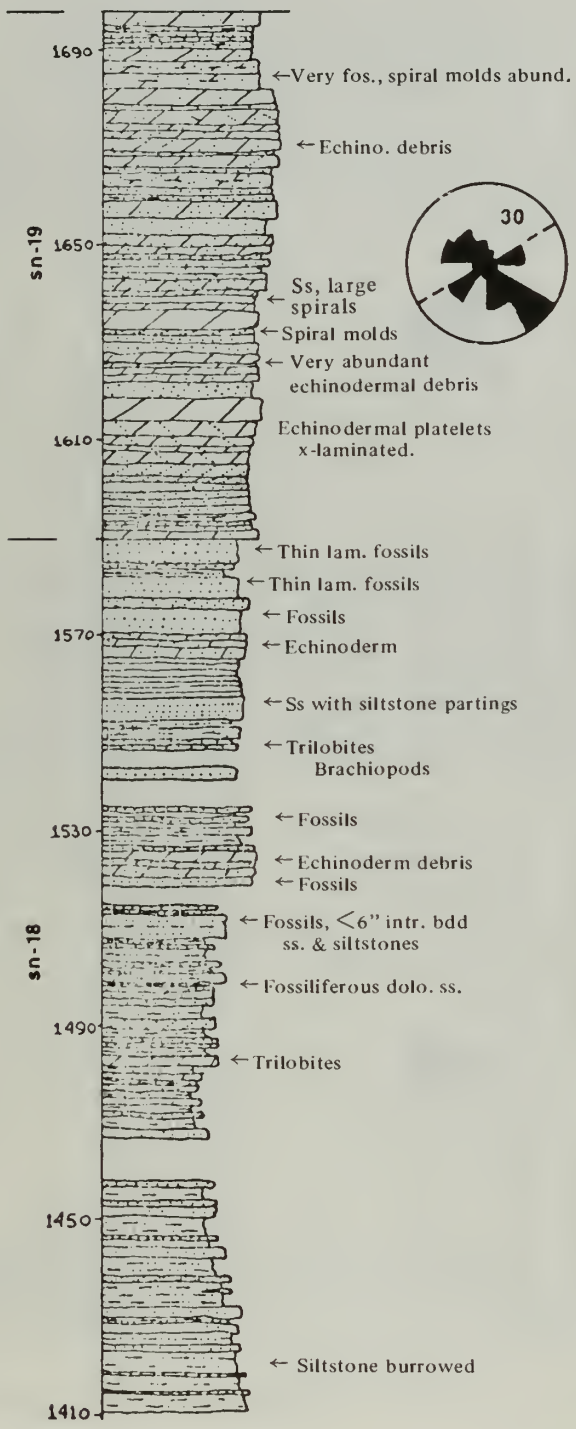
Figure 2 (continued).

by postdepositional tectonic rotation that occurred about a vertical axis and was related to the strike-slip movement of the Sheephead fault, which apparently passes immediately north of the peak.

The dolomite sub-units of the lower and upper members show gradual thickening to the northwest. The dolomite sub-unit of the upper member in the south Salt Spring Hills is 20 ft thick, whereas in Mosaic Canyon, at the north end of the Panamint Range, the correlative sub-unit is 340 ft thick. The three dolomite sub-units of the lower member are recognizable in each of the sections measured (except in the Salt Spring Hills, where the lower member is absent). They show a progressive thickening from southeast to northwest. These thickness trends are demonstrated in Figure 4a and b where thickness of the dolomite sub-units is plotted against distance from a line placed approximately parallel to the strike of the paleocurrent indicated by current rose diagrams for the two members. This line is positioned south of the Salt Spring Hills, where the lower member of the Wood Canyon Formation is absent.

Both the author's data and those of Stewart (1970) indicate that the lower member of the Wood Canyon Formation gradually thins to the southeast and disappears in the area of



sn-19

Intbdd dol, sandy dol, & qtztic ss. Dol m- cg, med dk gry(N4) wthrs olive gry (5Y4/1) w/ elephant hide texture, contains calcitic platelets which wthr in relief often defining crude xlams, oolitic. Dol f – mg, med dk gry(N4) wthrs dk yelsh br(10YR4/2), sandy dol w/ ribbons & stringers as well as xlam of brnsh blk(5YR2/1) f –mg silica cemented ss. Dol v. commonly xlam 32–60' from base of unit, xlam small scale w/ cosets 6" & less, foresets v. tangential bottoms w/ truncated tops, some mod red(5R4/6) slty wavy lam 1/8". Intbds of 1–3', lt gry, horiz lam, calc cemented qtzt, often xlam in middle part of unit. Ss becomes more common and more dolomitic upward w/ trilobite sand cast fragments and spiral molds abundant. Cyclic bdd is present but not as well displayed as in the sections to the NW.

sn-18

Gentle slope forming predominantly yelsh tan & lt olive gry fissile sltst in 6'-thick units w/ some v. thin beds of f – vfg horiz lam qtzt similar to that in SN-17. Toward top of unit cg, micaceous, highly fossiliferous ss is more common. Burrows & trails, load casts, flute casts, & ripples(interference) in phyllitic looking ss. Top ½ to 1/3 of unit dol beds appear and ss dominates w/ lt gry fossiliferous qtztic ss intbdd w/ dol & dol ss. Dol contains calcitic platelets – echinodermal debris? Fossils associated w/ m – cg lt gray qtztic ss beds w/ xlam darker less well cemented dol ss tops. Trilobite fragments, brach molds, spiral molds, vertical tubes abundant.

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Figure 2 (continued).

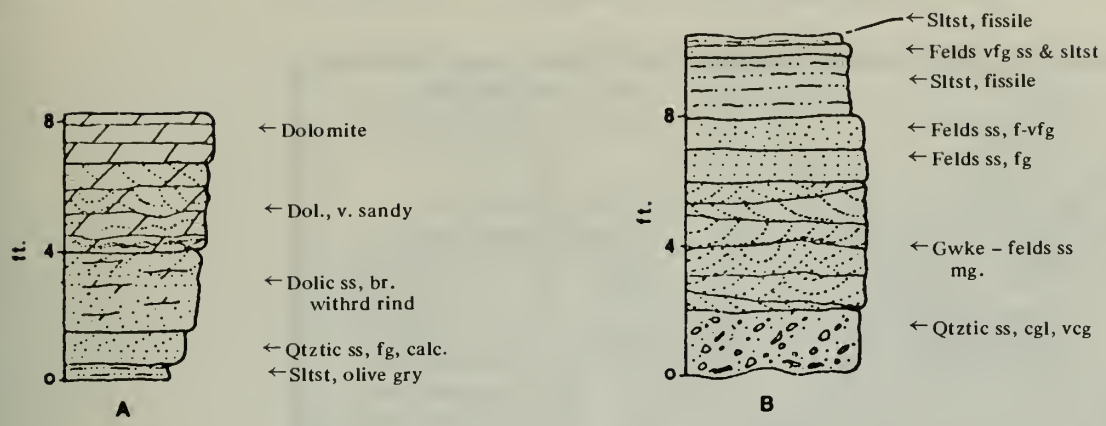


Figure 3. Cyclic bedding in: A. Upper member dolomite unit (sn-19); B. middle member cross-laminated sandstone and siltstones (sn-11 to 14). Scale 1 in. = 60 ft.

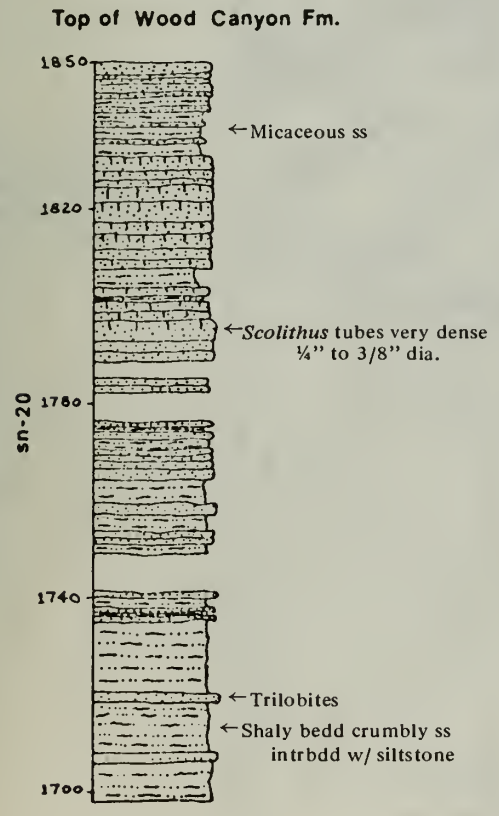
the Silurian Hills (Fig. 4b). In that area and to the southeast, conglomeratic arkose and arkosic sandstone of the middle member rest upon the Stirling Quartzite. Farther southeast, the Stirling disappears and the Wood Canyon correlative, the Tapeats Sandstone, lies directly upon older Precambrian gneiss and schist (Stewart, 1970, p. 39). This suggests a high to the southeast that is consistent with the paleocurrent indicators for the upper and lower members and is also compatible with the northwest thickening of the dolomite sub-units.

In contrast to the northwestward thickening of dolomite in the lower and upper members, no thickness trend is apparent in the middle member when it is considered in total. However, when the thickness of the arkosic conglomerate unit is plotted against distance southwest from an arbitrarily defined line which is parallel to the average strike of the paleocurrents indicated by the arkosic conglomerate rose diagrams, a marked thinning of sub-units to the southwest is obvious (Fig. 5b). Maximum diameters of pebbles measured in the arkosic conglomerate sub-unit at each section also decrease to the southwest (Fig. 5c).

PRELIMINARY INTERPRETATIONS

Bedding features and trace-fossil content, as well as current rose patterns mentioned above (Figs. 2, 3, 4a, 5a) suggest a tidal marine environment for the Wood Canyon Formation.

The vertical changes in paleocurrent directions demonstrable within single stratigraphic sections are consistent



Sn-20

Base of unit tan sltst intercalated w/ "crmbly" wthrng lt yelsh tan mg, v. micaceous ss in 2–5" beds. Also 9" & less beds of massive, f – mg, well cemented, qtztic ss, dk yelsh br(10YR4/2) wthrs to grysh br(5YR3/2); contains trilobite fragments. A less micaceous, mod red br wthrng, *Scolithus* bearing ss appears 1/3 the way up-section from base. Grnsh gry fissile sltst occurs nearer top of unit along w/ the tan sltst. Near top is 40', v. well cemented, dk yelsh orange(10YR6/6) qtzt wthrs mottled lt br(5YR5/6), v. dense *Scolithus*, beds 2–4' thick.
Top of Wood Canyon drawn at base of purplish red & red purple banded well cemented, vitreous qtzt.

Figure 2 (continued).

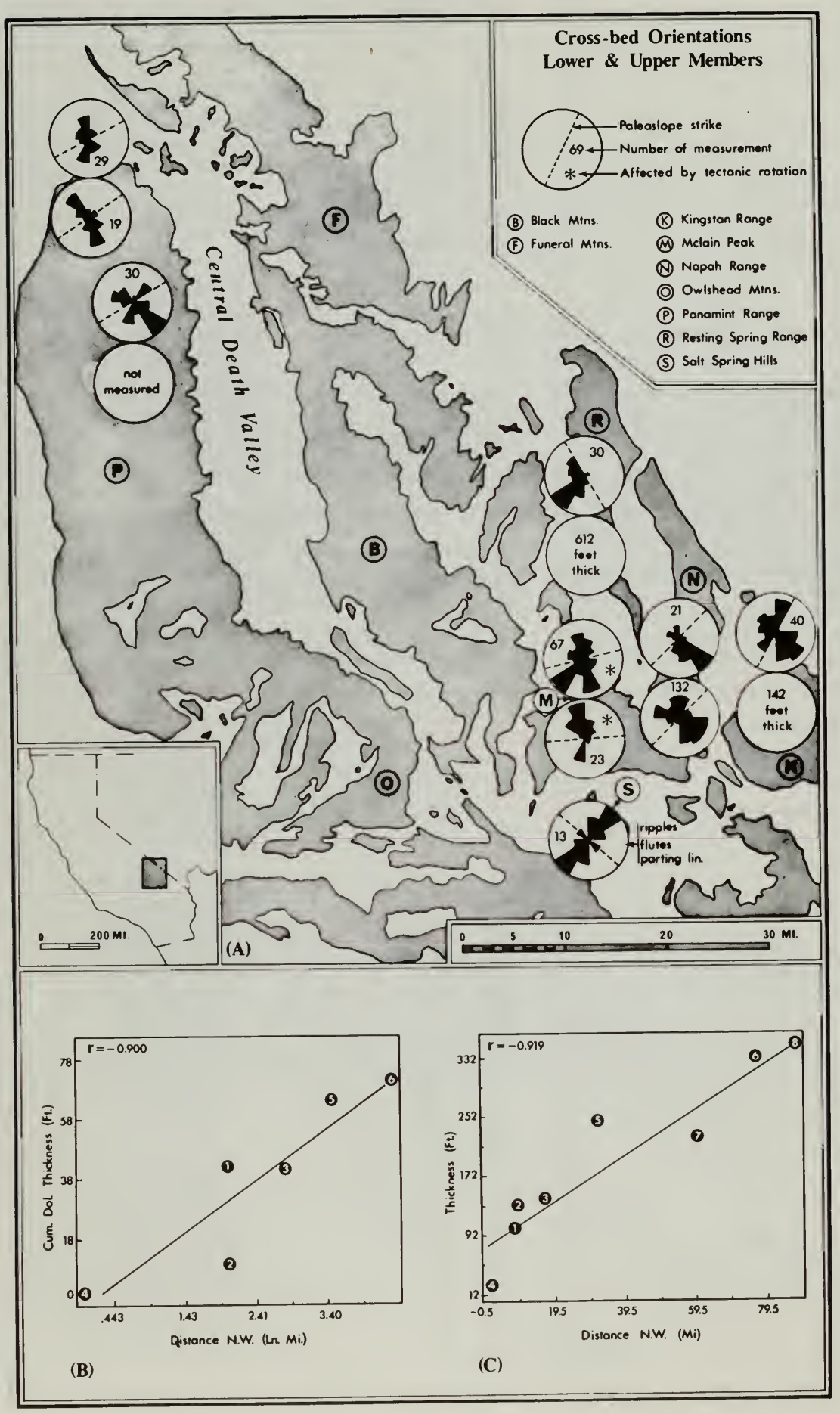


Figure 4. Cross-bed orientations of lower and upper members, Wood Canyon Formation.

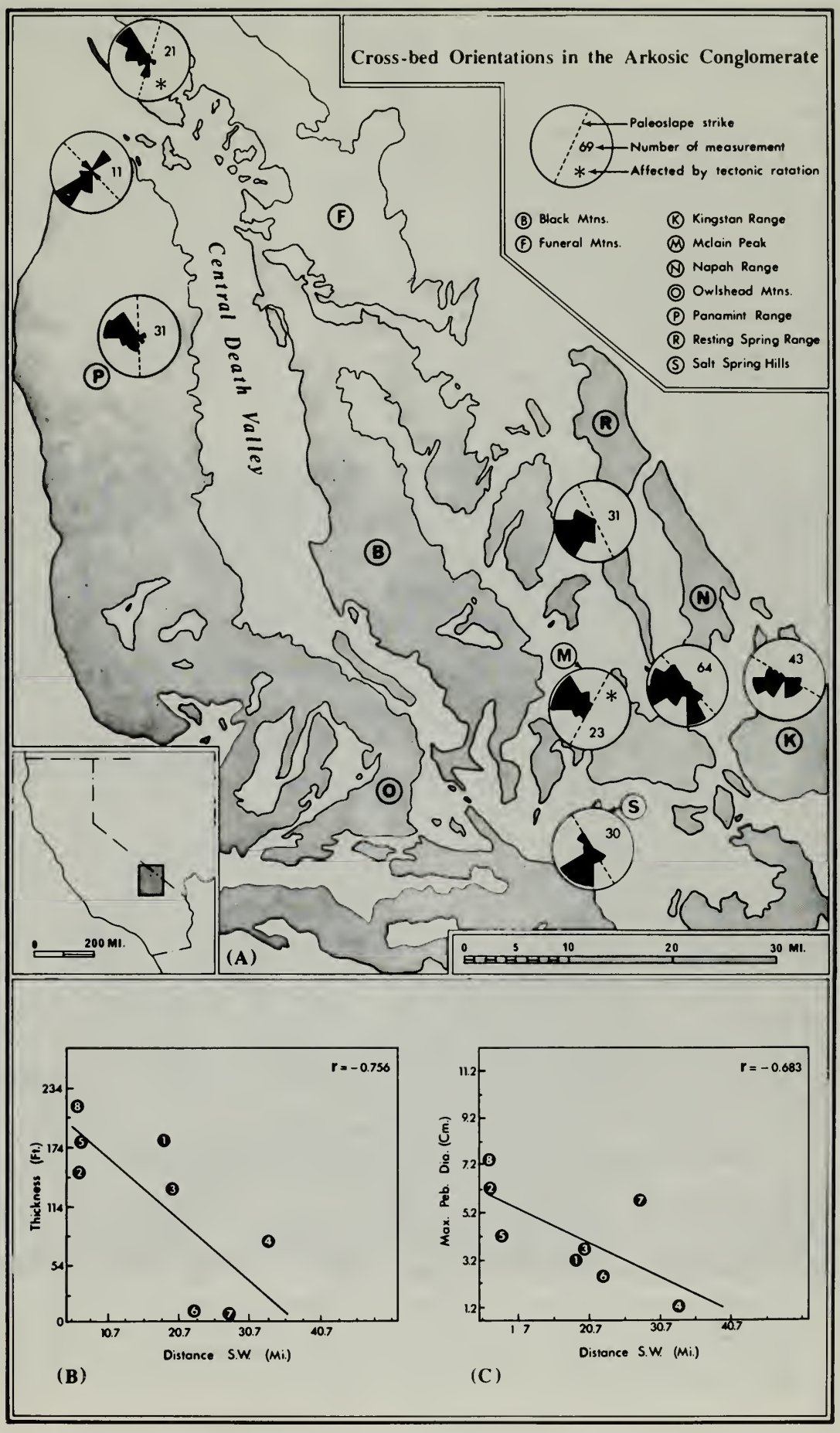


Figure 5. Cross-bed orientations in arkosic conglomerate, Wood Canyon Formation.

regionally and indicate a change in paleoslope during deposition of Wood Canyon strata. That the paleoslope was northwest during deposition of most of the formation is supported both by cross-lamination frequency diagrams and by the northwest increase of dolomite thickness and content in the upper and lower members. The northwest thickening of dolomite units, which is interpreted as a basinward increase in carbonate content, the existence of a southeastern high indicated by the gradual southeasterly disappearance of the lower member, and the paleocurrent data are consistent with a northwest paleoslope during the time of deposition of lower and upper Wood Canyon strata.

On the other hand, the southwestward thinning of the arkosic conglomerate, the decrease of maximum pebble diameter, also to the southwest, and predominantly unimodal paleocurrent frequency diagrams point to a southwest paleoslope during deposition of that coarse sediment sub-unit.

Thus, the data indicate a sharp change in transport direction from northwest to southwest at the base of the middle members with a gradual re-establishment up-section of the northwest trends shown in the lower member. This change in slope is intimately associated with the deposition of the arkosic conglomerate at the base of the middle member and is interpreted as indicating tectonic activity at that time. The arkosic sediments were probably derived from a granitic terrain to the east-northeast. The overlying and underlying beds may then well represent early stages of the Cordilleran miogeosynclinal development with paleoslope here to the northwest.

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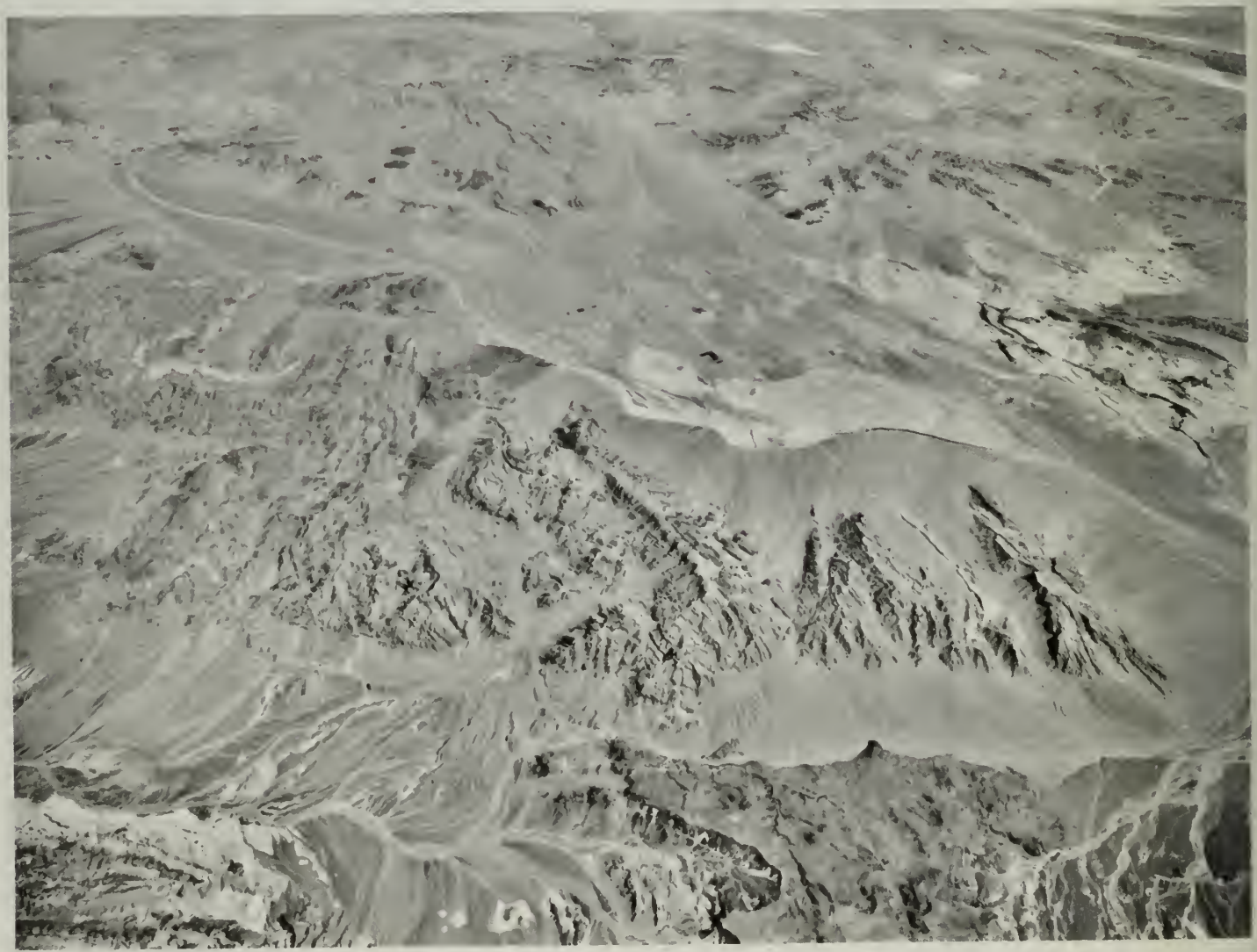

High-altitude oblique aerial photograph to northeast across southern Funeral Mountains taken from a point barely south of Ryan (lowermost center). Trace of Northern Death Valley-Furnace Creek fault zone extends across lower one-third of photograph. Rocks mostly Tertiary in age lie on near side of fault zone; late Precambrian and Paleozoic rocks lie beyond the fault, in Funeral Mountains. Rocks on left in the range are older and are moderately to strongly folded; rocks on right are younger and deformed mainly by eastward tilting. Tertiary sedimentary rocks lie at extreme right end of range. All rocks are repeated by abundant west-dipping normal faults. A major part of the Basin and Range province occupies rest of photograph. Ranges are dark-colored; basins contain pale-colored playa sediment. *Photo U.S. Air Force 374L 193, 6 September 1968; courtesy of the U.S. Geological Survey.*

Geologic Maps and Sections of a Strip from Pyramid Peak to the Southeast End of the Funeral Mountains, Ryan Quadrangle, California

James F. McAllister[1]

Work done partly in cooperation with the California Division of Mines and Geology.

Table 1. Explanation of map units. Maps are contiguous, aligned along A–A″. (Thickness is rounded to nearest 100 in maximum section.)

Qal	ALLUVIUM (Quaternary)
Qao	OLDER ALLUVIUM (Quaternary)
Qlds	LANDSLIDE MATERIAL (Quaternary)
QTf	FUNERAL FORMATION (Pleistocene? and Pliocene)
Tcl	CONTINENTAL CLASTIC ROCKS (Tertiary)
Tls	LACUSTRINE LIMESTONE (Tertiary)
Mp	PERDIDO FORMATION (Upper and Lower(?) Mississippian) 500 feet (top eroded)
Mt	TIN MOUNTAIN LIMESTONE (Lower Mississippian) 300 feet
Dl	LOST BURRO FORMATION (Upper and Middle Devonian) 2,500 feet
DSh	HIDDEN VALLEY DOLOMITE (Lower Devonian and Silurian) 1,400 feet
Oes	ELY SPRINGS DOLOMITE (Upper and Middle Ordovician) 500 feet
Oe	EUREKA QUARTZITE (Middle Ordovician) 400 feet
Op	POGONIP GROUP (Middle and Lower Ordovician) 2,200 feet
-Cn	NOPAH FORMATION (Upper Cambrian) 1,700 feet
	BONANZA KING FORMATION
-Cbb	BANDED MOUNTAIN MEMBER (Upper and Middle Cambrian) 2,400 feet
-Cbp	PAPOOSE LAKE MEMBER (Middle Cambrian) 1,200 feet
-Cc	CARRARA FORMATION (Middle and Lower Cambrian) 1,600 feet
Lower formations exposed in the Funeral Mountains, Ryan quadrangle	
	ZABRISKIE QUARTZITE (Lower Cambrian) 800 feet
	WOOD CANYON FORMATION (Lower Cambrian and Precambrian) 4,000 feet
	STIRLING QUARTZITE (Precambrian) 4,800 feet
	JOHNNIE FORMATION (Precambrian) 1,000 feet to concealed part

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[1] *U.S. Geological Survey, Menlo Park, California 94025.*

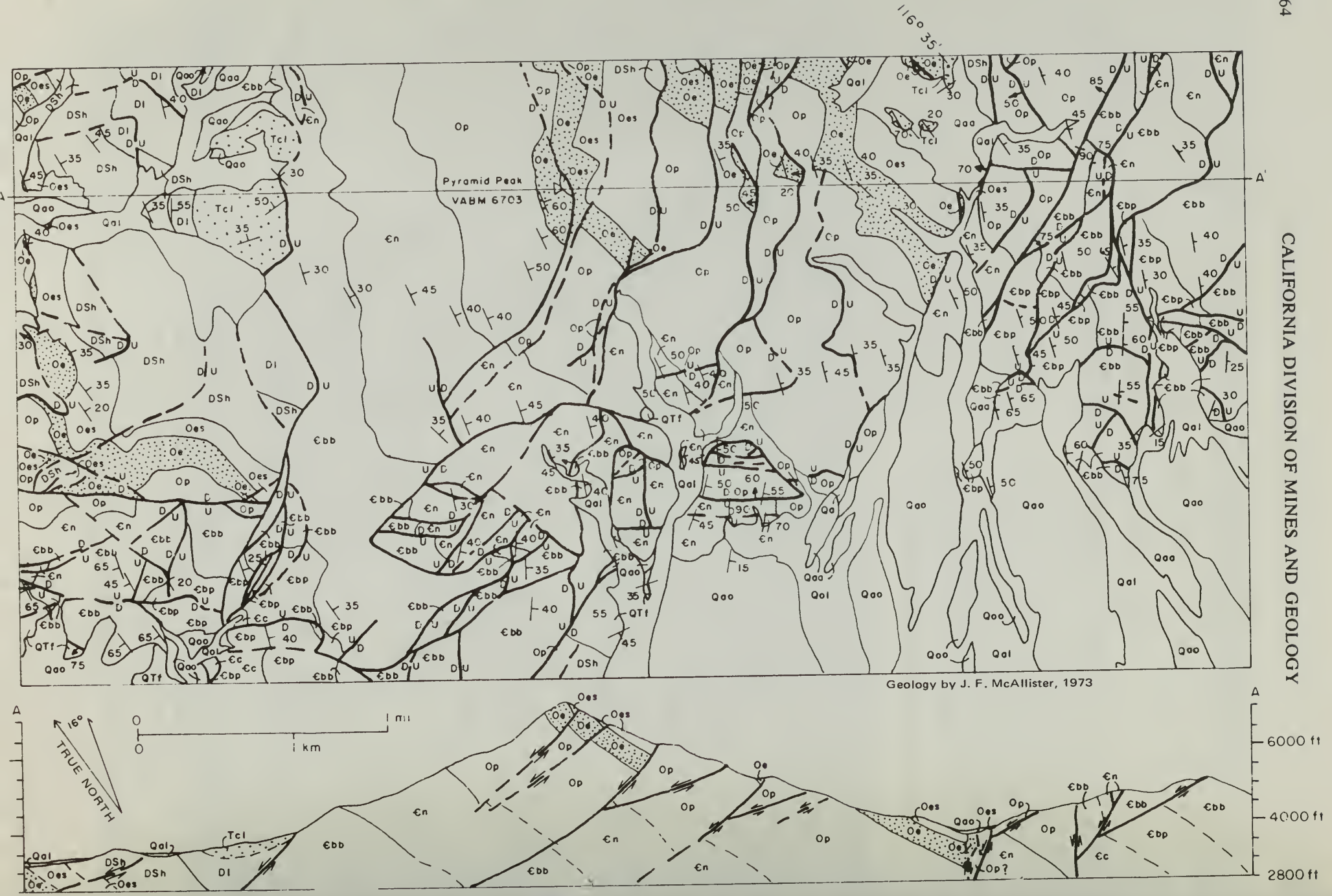

Figure 1. Geology at Pyramid Peak, Funeral Mountains. Explanation of map units in Table 1.

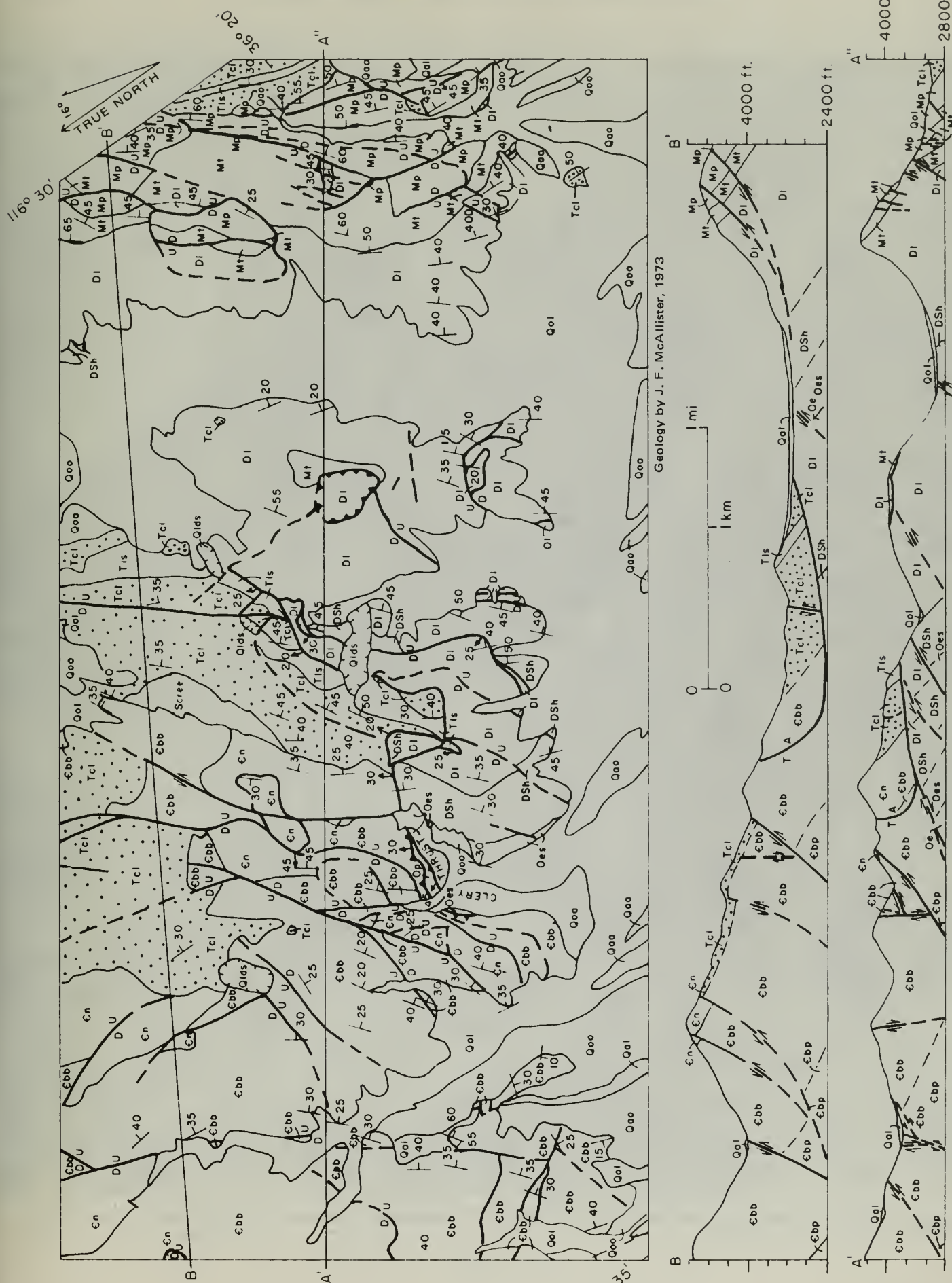

Figure 2. Geology at southeast end of Funeral Mountains. Explanation of map units in Table 1.

High-altitude imagery of southern and central Death Valley. *Imoge NASA ERTS 1125-17554 bond 7, 25 November 1972; copy produced by Drew P. Smith, Californio Division of Mines ond Geology.*

Geology of the Shoshone Volcanics, Death Valley Region, Eastern California

Richard Haefner[1]

GEOLOGIC SETTING

Within the Black Mountains block of the Death Valley region of eastern California is a terrane composed primarily of Cenozoic intrusive and extrusive rocks, which covers an area of about 1,300 sq km. The volcanic rocks of this terrane are predominantly rhyolitic but also include dacitic, andesitic, and basaltic units. They appear to be genetically related to the Cenozoic deformational features of the block (Noble and Wright, 1954; Wright and Troxel, 1971a, 1971b). The extrusive rocks that have been mapped thus far are mostly lava flows and air-fall tuffs, but ash-flow tuffs form much of the lower part of the volcanic sequence in the southern part of the field (L. A. Wright, personal commun.). Included within the rhyolitic rocks is an accumulation of lava flows and tuffs, as much as 900 m thick, to which the name "Shoshone Volcanics" is here applied and which is the subject of this summary report.

Although the field is incompletely explored, it appears to consist dominantly of acid intrusive and extrusive rocks; rocks of intermediate to basic composition are also present but subordinate.

Clastic sediments derived from the igneous rocks, together with evaporites, were deposited in local basins (Drewes, 1963). Some of these rocks contain commercial borate deposits. Two major sediment accumulations have been named Copper Canyon Formation (Curry, 1941; Drewes, 1963, p. 32) and Furnace Creek Formation (Noble, 1941, p. 956).

DESCRIPTION

The Shoshone Volcanics are exposed almost continuously in a belt about 26 km long, which extends northward (Fig. 1). They also are exposed in more westerly parts of the Greenwater Range and in the Black Mountains.

In the area of Figure 1, the Shoshone Volcanics overlie Cambrian sedimentary rocks, Cenozoic quartz monzonite plutons, and older acid volcanic rocks with an erosional unconformity. The contact between the two accumulations of volcanic rock shows marked angular discordance. Southwest of the area shown in Figure 1, the Shoshone Volcanics also overlie ash-flow tuffs and andesite lava flows. The eroded surface of the Shoshone Volcanics is in turn overlain by air-fall tuffs and lava flows of the Greenwater Volcanics. This contact commonly also shows angular discordance. As the dip direction of all three volcanic accumulations is generally eastward, the angular unconformities suggest continued eastward tilting of fault blocks during emplacement of the volcanic rocks. Dark colored lava flows of basaltic and andesitic composition overlie the Shoshone Volcanics in some places.

The Shoshone Volcanics are prominently layered, the layers producing a striped landscape colored in various shades of pink, yellow, and gray. A regular sequence of layers is obvious, even from a distance. A yellow layer is generally overlain successively by a gray layer, a pink layer, locally a second gray layer, and another yellow layer (Table 1). At many localities, three or four of these sequences are exposed in layer-cake

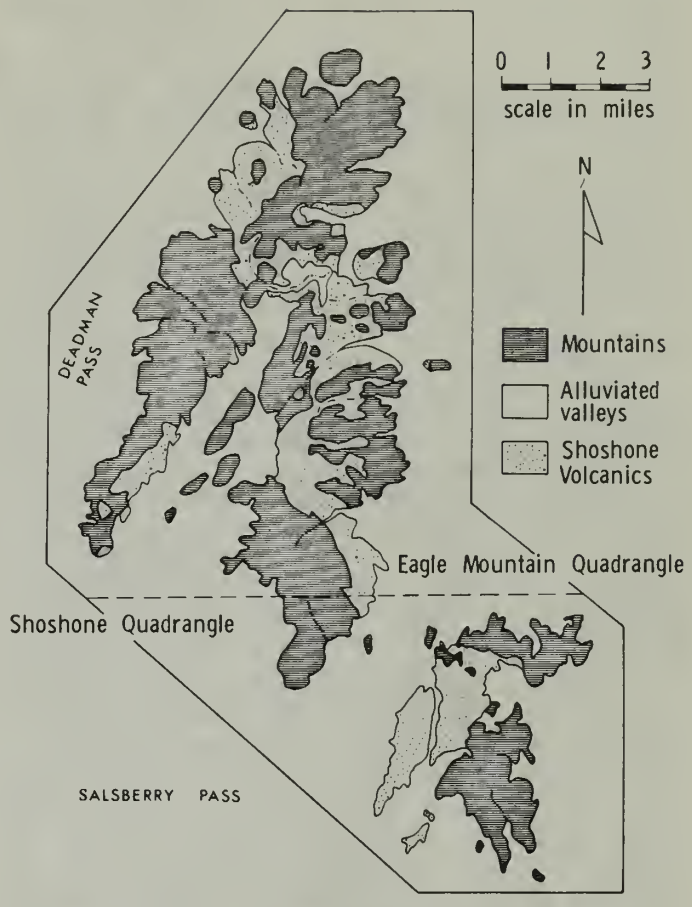

Figure 1. Distribution of the Shoshone Volcanics in southeast Greenwater Range (Eagle Mountain quadrangle) and Dublin Hills (Shoshone quadrangle).

fashion on a single slope, individual sequences being 45 to 120 m thick. The sequences on steeper slopes produce cliffs (gray and pink layers), alternating with benches (yellow layers).

As the volcanic rocks dip eastward, the best views of the layer sequences are from the west side of the mountains. The sequences in the Dublin Hills are among the best exposed (Chesterman, 1973) and are also easily accessible. These dominate the view eastward as one travels toward Salsberry Pass from Greenwater Valley.

CORRELATION AND AGE

The Shoshone Volcanics are equivalent to a part of Drewes's "older volcanics" in the Funeral Peak quadrangle to the west of the area shown in Figure 1 (Drewes, 1963). This unit is lithologically similar to the Shoshone Volcanics even to the zones within individual lava flows, and it also overlies eroded quartz monzonite plutons and underlies Greenwater Volcanics.

A tentative age for the Shoshone Volcanics may be inferred from isotope ages of rock units elsewhere in the volcanic field. Fleck (1970, p. 2810) reported seven K-Ar ages for

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"older volcanics" from the Funeral Peak quadrangle: three from Dante's View and four from Hidden Springs. The dates from Dante's View (6.32 ± 0.13, 6.34 ± 0.13, 6.49 ± 0.13) are clearly younger than those from Hidden Spring (8.02 ± 0.16, 7.60 ± 0.30, 7.77 ± 0.15, 8.02 ± 0.16). These yield an early Pliocene age for Hidden Spring and a middle Pliocene age for Dante's View, based on the time scale of Evernden and others (1964). Although the "older volcanics" of Drewes include more than the equivalent of the Shoshone Volcanics, the early to middle Pliocene dates fit the permissible interval allowed the Shoshone Volcanics by radiometric dating of the quartz monzonite plutons and the Greenwater Volcanics. The former have yielded middle to late Miocene K-Ar dates (Stern and others, 1966; Armstrong, 1966). The Greenwater Volcanics, on the other hand, have yielded late Pliocene K-Ar dates (Fleck, 1970). The early to middle Pliocene age for the Shoshone Volcanics thus indicated may be refined when more detailed correlations of volcanic units are made, and when the age of the volcanic rocks underlying the Shoshone Volcanics is better known.

PETROGRAPHY

The Shoshone Volcanics consist primarily of four lithotypes: (1) gray perlitic vitrophyre, (2) red-brown to pinkish-gray felsophyre, (3) yellow felsophyre, and (4) yellow devitrified tuff (Table 1).

Phenocrysts of the Vitrophyre and Felsophyre. Phenocrysts comprise 3 to 20 percent of the total rock volume (Table 2); the remainder is glass or cryptocrystalline groundmass. Some of the layer sequences are crystal rich, others are crystal poor (Haefner, 1972). The Dublin Hills sequences are crystal poor. Crystal-rich sequences overlie crystal-poor sequences in the vicinity of Brown Peak.

Plagioclase, the only feldspar, forms the largest (4 mm) and most abundant phenocrysts. Most crystals are andesine, but the composition ranges from oligoclase to labradorite.

Hornblende forms euhedral prisms to 1.5 mm long. It occurs as ordinary hornblende (pleochroic in green and brown), and as the varieties variously known as oxyhornblende, lamprobolite, or basaltic hornblende (pleochroic in red, green, and brown). Crystals of the latter variety are commonly rimmed with opaque minerals. Oxyhornblende occurs only in the pink felsophyre; ordinary hornblende occurs in the vitrophyre and yellow felsophyre.

TABLE 1. TERMINOLOGY AND INTERPRETATION OF LAYERING IN SHOSHONE VOLCANICS

Color bands	Topographic expression (in steep areas)	Zones*	Limits of eruptive unit	
Yellow	Slope	Lower altered vitrophyre zone		
		Tuff zone		
		Upper altered vitrophyre zone		
Gray	Cliff	Upper vitrophyre zone (not always present)	Single lava flow	Eruptive unit
Pink	Cliff	Red felsophyre zone		
Gray	Cliff	Lower vitrophyre zone		
Yellow	Slope	Lower altered vitrophyre zone		
		Tuff zone	Air-fall tuff	
		Upper altered vitrophyre zone		

* Zones are mappable units, drawn in this column in their average proportional thicknesses. Altered vitrophyre zones consist predominantly of yellow felsophyre, which at a distance may be difficult to distinguish from the tuff zone.

Biotite occurs as euhedra to 1.5 mm in maximum dimension. It, too, occurs as two varieties: ordinary biotite (pleochroic in green and brown), and a type which, by analogy with hornblende, may be called oxybiotite (pleochroic in red, green, and brown). Oxybiotite is commonly rimmed with opaques, and it occurs exclusively with oxyhornblende in the pink felsophyre.

Opaque minerals, largely magnetite, occur as small anhedral to euhedral grains from dust size to 0.3 mm, and as larger masses partly to completely replacing oxyhornblende and oxybiotite. Orthopyroxene and clinopyroxene form euhedral phenocrysts that may be as large as the hornblende and biotite but are much less common.

A xenocrystic character is inferred for some grains that are rounded or embayed. Quartz grains as much as 1 mm in diameter occur in a few thin sections and are consistently rounded or embayed. Species that occur as phenocrysts also commonly show such outlines.

Groundmass of the Vitrophyre and Felsophyre. The glass groundmass of the vitrophyre commonly is gray to black. Brown glass occurs as a local variation that is mixed in swirls and patches with gray glass. The glass contains abundant pyroxene crystallites of several habits. Perlitic structure is nearly ubiquitous. In thin section, concentric fractures outline spheres with centers spaced at intervals of 0.5 to 1.0 mm and associated with a boxwork of planar fractures. Spherulites, 0.1 to 30 cm, occur in the glass but are confined to the several meters of vitrophyre along the margins of the red felsophyre (Fig. 2). Some spherulites are hollow and contain quartz crystals, agate, or opal.

The groundmass of the yellow felsophyre is cryptocrystalline. The presence of the alteration mineral jarosite is indicated by x-ray diffraction analysis. Magnetite commonly is absent. Relict perlitic fractures are locally present, indicating that the yellow felsophyre groundmass was originally glass, and that it acquired its present crystallinity through devitrification.

The groundmass of the red felsophyre is cryptocrystalline. The red to pink color, which is particularly conspicuous around magnetite microphenocrysts and around the opacitized rims of oxyhornblende and oxybiotite, is produced by abundant patches of hematite. In contrast to the vitrophyre and yellow felsophyre, the red felsophyre is vesicular; small vesicles to 2 mm are nearly ubiquitous, but large vesicles to 1 cm also occur, most of them adjacent to the vitrophyre (Fig. 3).

Devitrified Yellow Tuff. The same phenocryst species as found in the vitrophyre and felsophyre are found in the tuff matrix. The matrix has been devitrified and is now cryptocrystalline, with only rare traces of the original glass shards. Magnetite appears to be absent in the matrix, whereas jarosite is present. Essential pumice lapilli (devitrified) and accidental lapilli and blocks of volcanic rocks are the most common clasts; clasts of sedimentary and metamorphic rocks are rare. Vitrophyre clasts are absent nearly everywhere, but yellow and red felsophyre clasts are common.

Two localities have been found where the tuff is not devitrified. The most accessible of these is the lowest tuff

TABLE 2. APPROXIMATE MODAL COMPOSITIONS, BY CONVENTIONAL POINT COUNTING

	Crystal-poor eruptive unit* (4,200 points) (%)	Crystal-rich eruptive unit† (2,000 points) (%)
Groundmass (glass)	97.75	81.8
Plagioclase	2.02	11.7
Biotite	0.12	0.6
Hornblende	0.07	4.7
Opaques	0.02	1.2
Augite/hypersthene	0.02	tr.
Total	100.00	100.0

* Dublin Hills, average of 21 vitrophyre thin sections from one lava flow.
† West of Brown Peak, average of 10 vitrophyre thin sections from one lava flow.

exposed on the west side of the Dublin Hills where vitrophyre lapilli and glass shards are abundant, and jarosite does not occur in the matrix.

Paleomagnetic Properties of the Rocks. Ninety-six determinations of paleomagnetic direction were made on vitrophyre and red felsophyre of the layer sequences only because yellow felsophyre and tuff commonly are not magnetically susceptible. In each case, strike and plunge of the remanant magnetization was determined. As the rocks are relatively young, the orientation of the remanant magnetization should approximately parallel the magnetic direction of the Earth's present field, after correcting for postcooling deformation.

All the layer sequences examined exhibit normal polarity, except for specimens from exposures of one layer sequence, which are reversed (NW¼ sec. 3, T. 22 N., R. 5 E.).

All the vitrophyre samples, including those from the reversed sequence, have directions of remanant magnetization which parallel that predicted from the present field (Fig. 4). However, among specimens from the red felsophyre zone, fewer than half have remanant directions that parallel the predicted direction; the remainder have remanant directions with widely varying strikes and plunges. It appears that only the vitrophyre can be relied upon to be a faithful recorder of the Earth's magnetic field at the time of emplacement.

INTERPRETATION OF THE LAYER SEQUENCES

Types of Deposits. Within each yellow layer is a deposit, about 9 m thick, of well-stratified tuff. The remainder of the layer is yellow felsophyre (Table 1). The tuffs, which consist of interbedded lapilli-rich and lapilli-poor strata, are interpreted as air-fall tuffs. The tuff deposits divide the yellow color bands into additional mappable units, or zones (Table 1).

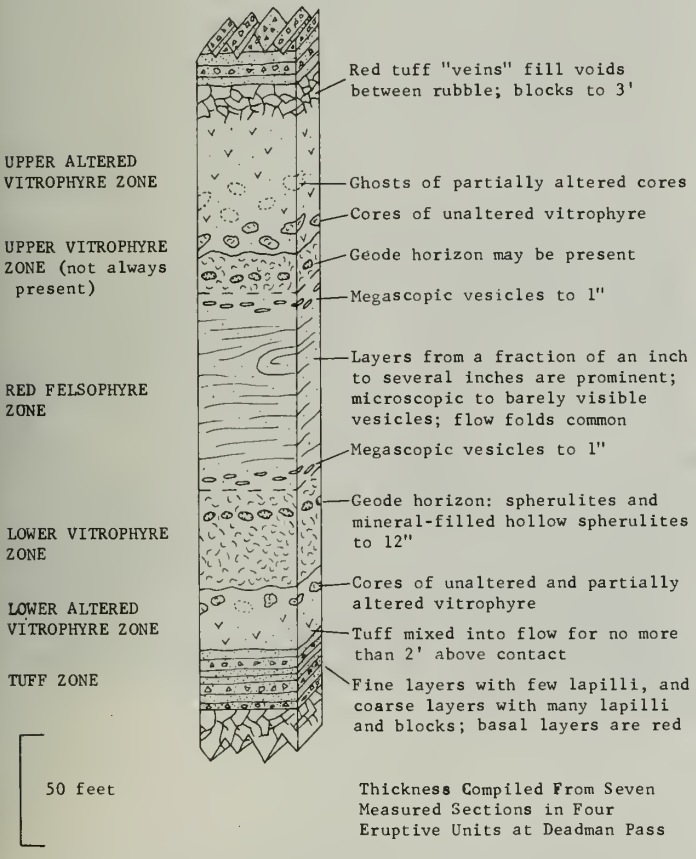

Red tuff "veins" fill voids between rubble; blocks to 3'
UPPER ALTERED VITROPHYRE ZONE
Ghosts of partially altered cores
Cores of unaltered vitrophyre
UPPER VITROPHYRE ZONE (not always present)
Geode horizon may be present
Megascopic vesicles to 1"
Layers from a fraction of an inch to several inches are prominent; microscopic to barely visible vesicles; flow folds common
RED FELSOPHYRE ZONE
Megascopic vesicles to 1"
Geode horizon: spherulites and mineral-filled hollow spherulites to 12"
LOWER VITROPHYRE ZONE
Cores of unaltered and partially altered vitrophyre
LOWER ALTERED VITROPHYRE ZONE
Tuff mixed into flow for no more than 2' above contact
TUFF ZONE
Fine layers with few lapilli, and coarse layers with many lapilli and blocks; basal layers are red
50 feet
Thickness Compiled From Seven Measured Sections in Four Eruptive Units at Deadman Pass

Figure 2. Features of typical eruptive unit in Shoshone Volcanics.

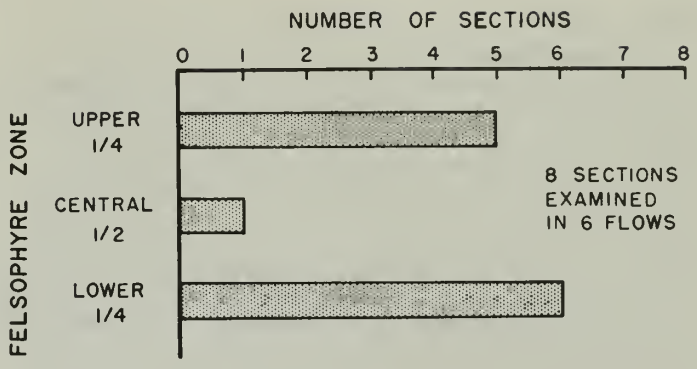


Figure 3. Distribution of megascopic vesicles in red felsophyre.

The sequence of zones between tuff deposits is interpreted as a single rhyolite lava flow (Fig. 2; Haefner, 1969). Glass shards, fiamme, and other characteristics of ash-flow deposits are absent in these zones. In addition, laminae in the red felsophyre zone are locally warped into folds, indicating that the magma flowed as a coherent mass. That this sequence of zones constitutes only one lava flow is evidenced by exposures of flow margins. Four flow margins are exposed in the Dublin Hills; at such localities, the red felsophyre constitutes a core, around which are wrapped the upper vitrophyre and yellow felsophyre zones to become the lower vitrophyre and yellow felsophyre zones (Fig. 5).

Alteration. The yellow felsophyre is interpreted as altered vitrophyre (Haefner, 1969, 1973). Evidence for this inference is the presence of relict perlitic fractures and the observation that yellow felsophyre merges with unaltered vitrophyre inward from both the upper and lower flow surfaces. Thus, a lava flow appears to consist essentially of a red felsophyre core encased in a vitrophyre sheath; the outer portions of the sheath have been altered, as have adjacent tuff deposits. Alteration is expressed as devitrification, the growth of jarosite, and the disappearance of magnetite.

Altered acid volcanic rocks are usually distributed about centers of fumarolic or hydrothermal activity. The Shoshone lava flows are unusual in that the altered rock occurs as a sheet-like body at the glassy top and at the glassy base of each lava flow. However, such alteration may not be unique to Death Valley; there are at least two localities in the Soviety Union from which apparently similar altered lava flows of acid composition have been described (Nasedkin, 1963).

A hypothesis for the origin of this alteration in the Shoshone Volcanics, which involves the interaction of magmatic with meteoric volatiles during the cooling of the flows, has been presented elsewhere (Haefner, 1969, 1973).

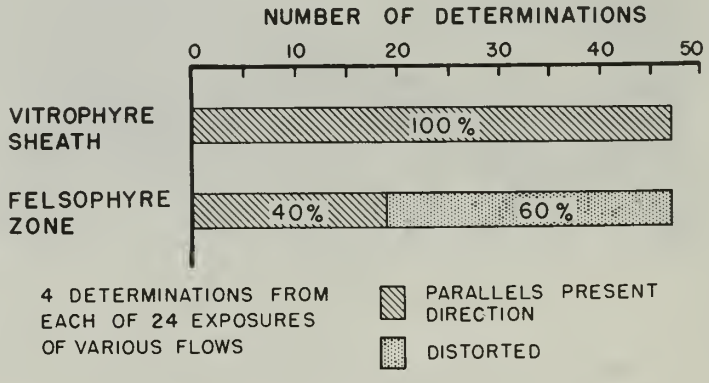


Figure 4. Determination of paleomagnetic direction.

EMPLACEMENT OF A TYPICAL ERUPTIVE UNIT

Crust of the Active Lava Flow. At the top of each lava flow that is overlain by a tuff deposit, veinlets of tuff extend downward into the flow, outlining individual blocks of yellow felsophyre (altered vitrophyre); locally, the blocks are pumiceous. These blocks apparently constituted a rubble crust formed by the chilling and breaking of lava on the upper surface of the active flow. The crust, about 4.5 m thick, composes only 3 to 9 percent of the total flow thickness; furthermore, crustal thickness does not depend on the thickness of the lava flow (Fig. 6).

Evidence of a rubble crust at the base of the lava flow would be generally difficult to detect, because tuff veinlets do not outline the rubble blocks. Nevertheless, ghostly remnants of angular blocks in the yellow felsophyre (altered vitrophyre) are observable at some localities and apparently originated as rubble. The lower crust is relatively thin, and massive, unaltered vitrophyre composes most of the part of the flow that underlies the red felsophyre zone (Fig. 2). If the lower crust was about as thick as the upper crust, then 80 to 95 percent of the lava flow remained liquid while the flow was active.

This large proportion of liquid magma contrasts with that inferred for many other acid lava flows, which are block lavas. These contained a much higher proportion of rubble during their active stage than do the flows of the Shoshone Volcanics. The overlying flows of the Greenwater Volcanics, exposed near Brown Peak, are block lavas. One measured section shows a flow to have been 70 percent rubble and only 30 percent liquid during its active stage.

Formation of Massive Vitrophyre. Adjacent to the rubble crusts are thick layers of massive gray vitrophyre and massive yellow felsophyre (altered vitrophyre). Together they form a sheath that encases the red felsophyre core of the flow. Patches of gray vitrophyre within the massive altered vitrophyre resemble rubble crust (Fig. 2), but they are not, because flow laminae or layers of spherulites can be traced across the patches into the adjacent vitrophyre, thus demonstrating that the patches are not rotated blocks.

The massive vitrophyre, fresh and altered, represents lava which must have chilled to a glassy rock only after movement of the lava ceased. Otherwise, the chilled glassy rock would have been broken and become part of the rubble crust. This chilling took place at least in part after the flow margin stopped advancing, because the massive vitrophyre sheath extends to the flow margin (Fig. 5). The chilling to form vitrophyre could

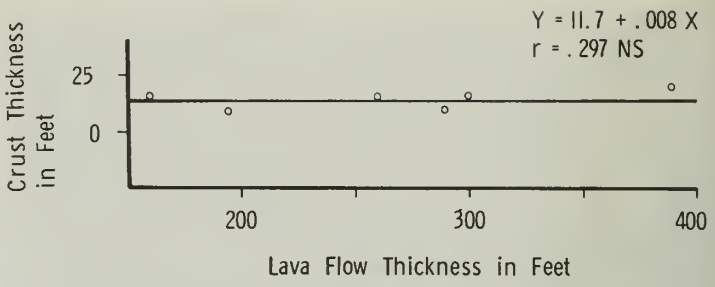


Figure 6. Thickness of upper crust of rubble in six lava flows of Shoshone Volcanics.

have begun in less mobile parts of the flow during the active stage, particularly in portions of the flow near the vent (Fig. 7).

Formation of the Felsophyre Core. An emplaced and cooling lava flow consisted of a molten core surrounded by a vitrophyre sheath (Fig. 7). The layers of massive vitrophyre must have continued to thicken as cooling proceeded, until conditions favored creation of features that characterize the lava flow core, namely, vesicularity, hematite stain, oxyhornblende and oxybiotite, and primary cryptocrystallinity, as contrasted with crystallinity caused by devitrification.

Except for local pumiceous rock in the rubble crust, only the felsophyre core is vesicular. This indicates that the lava, instead of degassing completely when it reached the Earth's surface, retained an appreciable quantity of gas in solution until after the flow was emplaced and had cooled for some time.

Lava flows of basic composition are generally devoid of evidence of delayed vesiculation, probably because they have a lower melt viscosity. The proportion of acid lava flows that display evidence of delayed vesiculation, is unknown, because detailed descriptions of the interiors of acid lava flows are rare. The delayed vesiculation probably is responsible for the other characteristic features of the red felsophyre, based on the following evidence: On heating in air to about 750°C, ordinary hornblende changes to oxyhornblende (Belovsky, 1891; Kozu and others, 1927; Barnes, 1930). This process is one of oxidation and is accompanied by resorption around the crystal margins, the products generally including iron oxides and pyroxene (Deer and others, 1962).

The conversion of ordinary hornblende to oxyhornblende and of biotite to oxybiotite in the flow interior probably occurred with vesiculation, the exsolved gases providing the

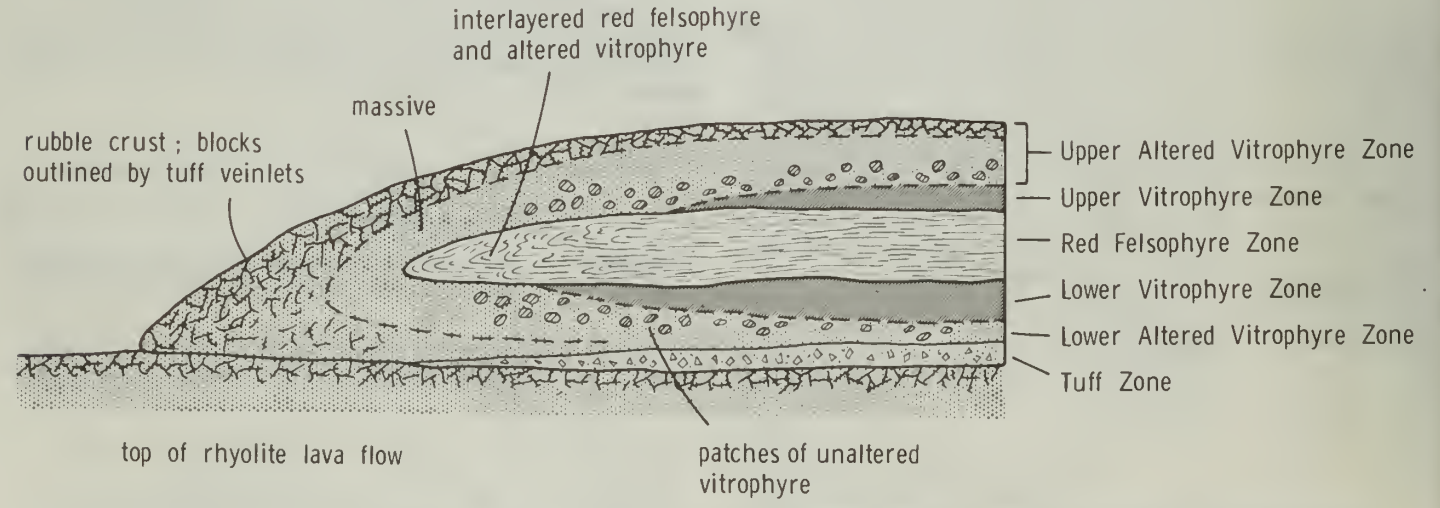


Figure 5. Cross-section sketch of lava flow margin in exposure of Shoshone Volcanics near Miller Spring. Flow margin overlain by another eruptive unit of Shoshone Volcanics. Not drawn to scale; length of section approximately 250 ft.

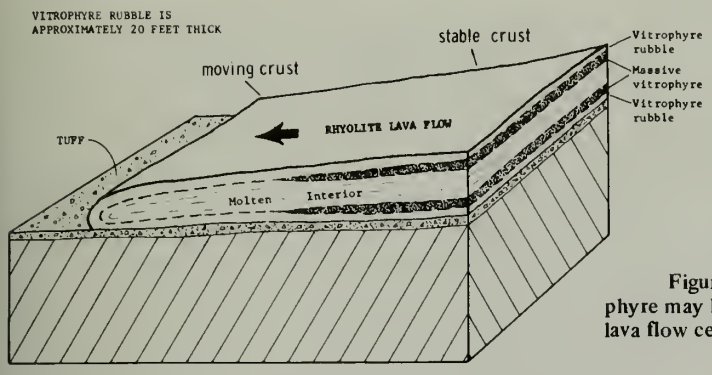


Figure 7. Interpretation of emplacement of an eruptive unit. Massive vitrophyre may have formed in stable crust areas of active lava flow but also formed after lava flow ceased advancing. Not drawn to scale.

necessary oxidizing environment. Because hematite, the colorant of the lava flow core, is concentrated around oxyhornblende and oxybiotite crystals, it probably formed as a resorption product.

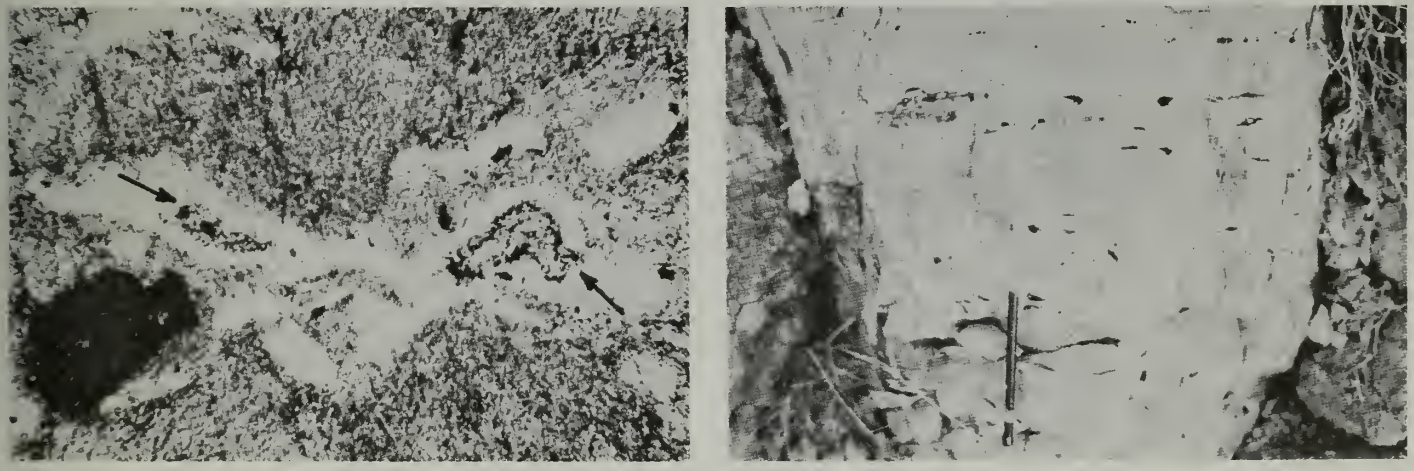
Because oxidation is an exothermic process, additional heat probably was generated in the molten lava flow core upon vesiculation. The oxidation of hornblende, experimentally, suggests that the temperature of the molten core was at least 750°C when the oxidation began. The additional heat provided by the oxidation apparently was sufficient to maintain the vesiculating molten core in a liquid state long enough for ions to migrate through the miscous melt and organize themselves as minute crystals. Thus the cryptocrystallinity of the red felsophyre groundmass also appears to have been favored by the delayed vesiculation.

Vesicle Collapse in the Felsophyre Core. The walls of vesicles are more coarsely crystalline than the rest of the groundmass. Streaks of similar material, observed in thin section, demonstrate the partial to complete collapse of vesicles (Fig. 8). Collapsing may have occurred when exsolving gases became depleted in the melt, or when the falling temperature caused a reduction in gas pressure to the point where it could no longer support the vesicle walls. The relatively large vesicles, which occur at the top and base of the red felsophyre zone, may have become "frozen" in highly viscous lava close to the chilled vitrophyre.

The peculiar paleomagnetic properties of the red felsophyre may be related to vesicle collapse. About 40 percent of determinations on red felsophyre yield magnetic directions that parallel the predicted directions of the rocks. The red felsophyre, even though it has been oxidized by exsolved gases, evidently can be a faithful recorder of the Earth's magnetic field, just as the vitrophyre can. Thus, the remaining determinations that do not yield the predicted magnetic direction seem best explained by mechanical rotation of mineral grains perhaps related to vesicle collapse. No other rotation is apparent as the paleomagnetism of the massive vitrophyre, particularly in the flow margins, demonstrates that the Curie point in the outer part of the lava flow was crossed only after the lava flow had stopped advancing. If this interpretation is correct, then the collapse of vesicles in the red felsophyre took place below 580°C, the measured Curie point of the rock.

Mechanism of Lava Flow Advance. Since freshly fallen tuff is notoriously susceptible to erosion but is conformable with the overlying flows, the lava flows must have moved over the tuff immediately after the tuff was deposited. Therefore, tuff is taken to represent deposits of explosions that cleared the vent just prior to a flow effusion.

The tuff deposits, which were unconsolidated lapilli and ash at the times the lava flows moved over them, are virtually undisturbed. Typically, the tuff fragments that are mixed into the overlying lava flow rock occur within 10 cm of the contact. In addition, the surfaces of the tuff layers apparently are unscoured by the overlying flows. Individual layers at the tops



A B

Figure 8. A. Photomicrograph of collapsed vesicles in red felsophyre (crossed nichols). Coarsely crystalline walls of vesicles show as light streaks. Arrows point to tridymite, which filled in partially collapsed vesicles. Light streak above tridymite patch on right is completely collapsed vesicle, without tridymite filling. Length of photo, 3 mm. B. Large vesicles in red felsophyre near bottom of zone. Vesicles are lined with druses of silica minerals.

of tuff deposits can be traced for nearly 1 km. Therefore, a typical lava flow advanced without disturbing the underlying loose tuff. The nature of the flow margins and the existence of rubble crust at the base of the typical flow indicate that it advanced with a simple rolling tractor-tread motion, the crust that formed at the upper surface being dragged beneath the flow as it advanced (Fig. 7).

Although the areal extent of individual lava flows is obscured by abundant faults, individual flows have been traced continuously for as much as 5.5 km. A lava flow in the Dublin Hills that contains distinctive fragments of basaltic rock may be correlative with a similar lava flow exposed as far as 19.3 km north of the Dublin Hills.

Considered as a group, the lava flows of the Shoshone Volcanics appear to have filled the valleys. A map of the area near Brown Peak in the Eagle Mountain quadrangle, for example, shows flows accumulated which filled an ancient valley that possessed a relief of about 300 m (Haefner, 1972, Pl. 2).

The tractor-tread motion, the large proportion of liquid lava in the active flows, and the apparent extensiveness of some lava flows suggest a high fluidity for the Shoshone lava flows. This might be an effect of high initial temperature (Gibbon, 1964), but the presence of phenocrysts of the lava suggests a limited temperature range comparable to the range of the relatively viscous lava assumed for the porphyritic flows of block rhyolite. Thus, apparently high fluidity of the Shoshone rhyolite lava flows is more likely an effect of a high content of dissolved volatiles than of a high initial temperature, the volatiles being largely retained in solution during the advance of the flows.

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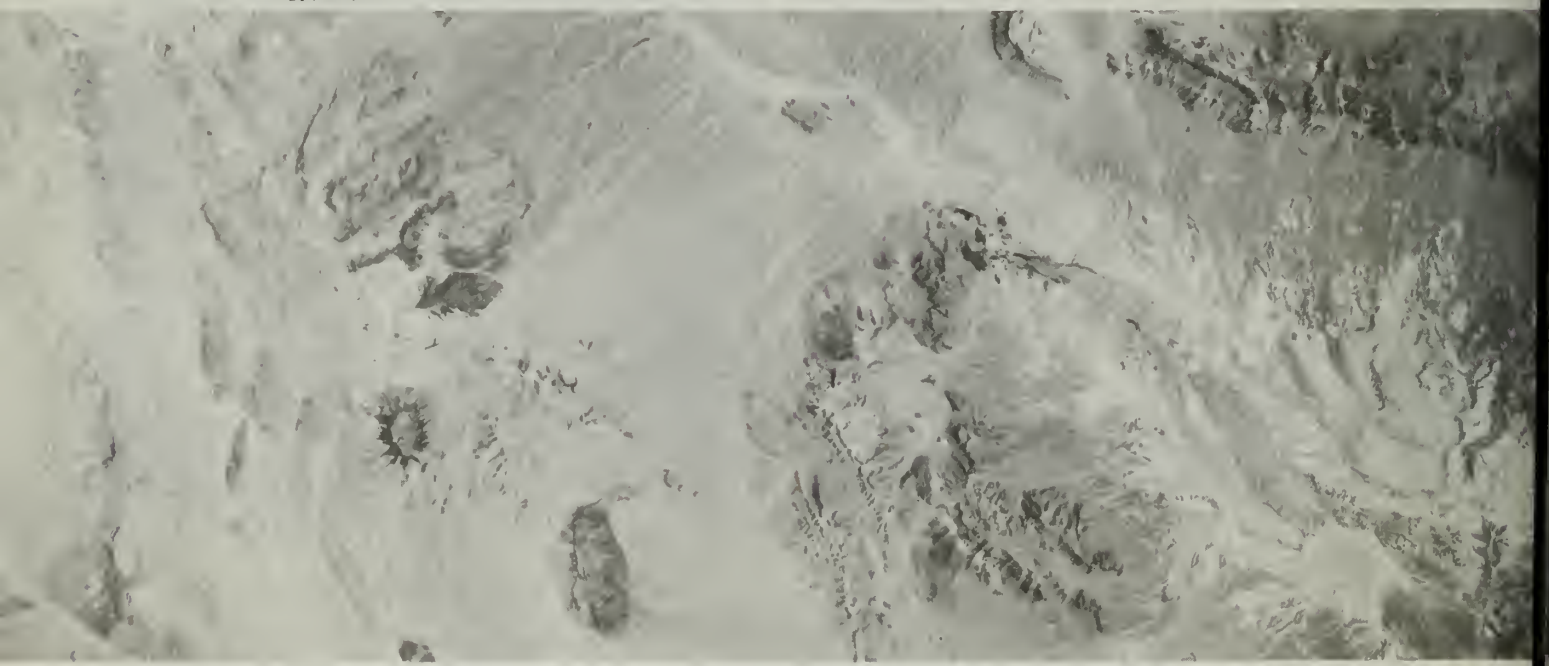
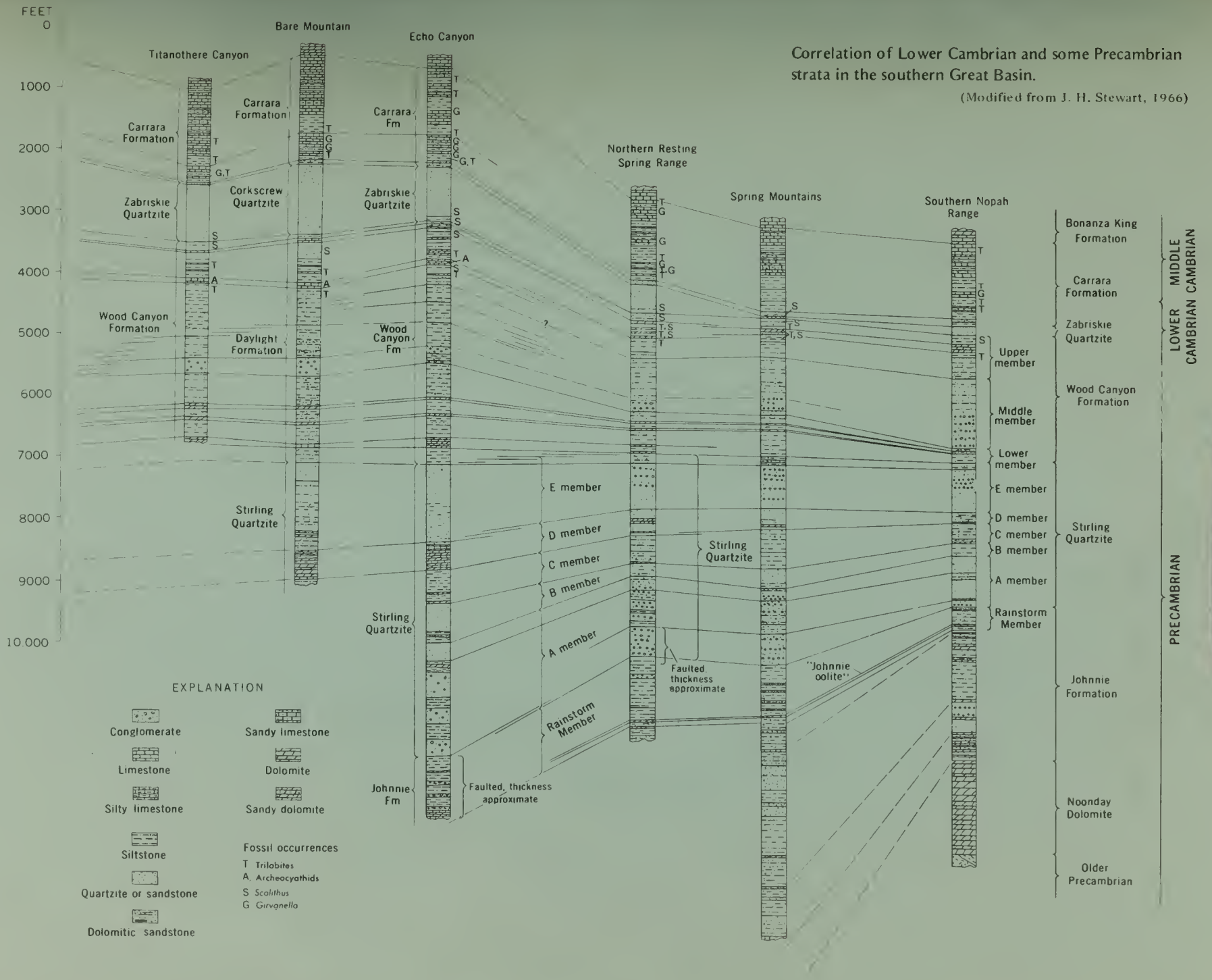


High-altitude vertical aerial photograph near southeastern end of Greenwater Range. Pale-colored rocks in left of center of photograph are marine Tertiary volcanic rocks described by Haefner (see preceding article). The rocks also crop out on northwest (left) side of Dublin Hills (right of center). Amargosa River flows south across photograph. Pale-colored area on right is underlain by flat-lying lake sediments deposited during an interval when the flow of the Amargosa River was blocked. *Part of photo U.S. Air Force 374V 197, 6 September 1968; courtesy of the U.S. Geological Survey.*



Correlation of Lower Cambrian and some Precambrian strata in the southern Great Basin.

(Modified from J. H. Stewart, 1966)

Back cover. High-altitude vertical photograph near south end of Death Valley. North is toward top of photograph. Main mass of Avawotz Mountains occupies lower right of photograph, partly covered by snow. Trace of north branch of Garlock fault forms obvious straight north boundary of part of Avawotz Mountains and hills to west. Southern Death Valley fault zone extends northwest to upper left corner of photograph from Avawatz Mountains and Garlock fault in right center of photograph. Small unnamed northwest-trending hills contain several branches of the Southern Death Valley fault zone; youngest branch bounds northeast side of hills and continues farther northwest in floor of valley.

Braided channels are part of Amargosa River which flows west around southern end of Saddle Peak Hills (upper right) and Saratoga Hills (upper center) then north into central Death Valley. *Photo U.S. Air Force 744V 073, 29 November 1967; courtesy of the U.S. Geological Survey.*

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